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ABSTRACT

One of the series for enlisted Navy and Naval Reserve personnel, the training manual prepares students for advancement within their rating, aids in on-the-job training, or serves as a review source. Charts, graphs, tables, maps, and illustrations supplement detailed text covering atmospheric physics and world climatology, air masses, circulation, and frontal systems. Several chapters focus on forecasting upper air systems, surface systems, cloud conditions, precipitation, temperature, storms, icing, fog, and so forth. Tropical conditions are given speical treatment, and several chapters deal with weather briefing, flight forecasting, sea surface and ocean thermal structure forecasting, and other special forecasts. Three chapters examine meteorological equipment and maintenance, and a final chapter covers administration, training, and communications. (MDW)

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AEROGRAPHER'S MATE 1 & C

NAVAL EDUCATION & TRAINING COMMAND

RATE TRAINING MANUAL

NAVEDTRA 10362-B

CE 003 557

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PREFACE

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This Rate Training Manual is one of a series of training manuals prepared for enlisted personnel of the Navy and Naval Reserve who are studying for advancement in the Aerographer's Mate (AG) rating. As indicated by the title, the manual is based on the professional qualifications for the rates AG1 and AGC, as set forth in the Manual of Qualifications for Advancement, NavPers 18068 (Series).

Combined with the necessary practical experience and a thorough knowledge of the material contained in AG 3 & 2, NavTra 10363-D, completion of the correspondence course based on this manual will greatly assist the AG2 and AG1 in preparing for advancement examinations. This training manual should also be valuable as a review source for the AGC who is studying for advancement to AGCS and the AGCS who is studying for AGCM. In addition, its everyday use in on-the-job training is highly recommended.

This training manual was prepared by the Naval Education and Training Program Development Center, Pensacola, Florida, for the Chief of Naval Education and Training Support. Technical review of the manuscript was provided by personnel of the AG(B) School, NATTC Lakehurst, N. J. Technical review and assistance was also provided by personnel of the Naval Weather Service Command.

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THE UNITED STATES NAVY

GUARDIAN OF OUR COUNTRY

The United States Navy is responsible for maintaining control of the sea and is a ready force on watch at home and overseas, capable of strong action to preserve the peace or of instant offensive action to win in war.

It is upon the maintenance of this control that our country's glorious future depends; the United States Navy exists to make it so.

WE SERVE WITH HONOR

Tradition, valor, and victory are the Navy's heritage from the past. To these may be added dedication, discipline, and vigilance as the watchwords of the present and the future.

At home or on distant stations we serve with pride, confident in the respect of our country, our shipmates, and our families.

Our responsibilities sober us; our adversities strengthen us.

Service to God and Country is our special privilege. We serve with honor.

THE FUTURE OF THE NAVY

The Navy will always employ new weapons, new techniques, and greater power to protect and defend the United States on the sea, under the sea, and in the air.

Now and in the future, control of the sea gives the United States her greatest advantage for the maintenance of peace and for victory in war.

Mobility, surprise, dispersal, and offensive power are the keynotes of the new Navy. The roots of the Navy lie in a strong belief in the future, in continued dedication to our tasks, and in reflection on our heritage from the past.

Never have our opportunities and our responsibilities been greater.

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CHAPTER 1

AEROGRAPHER'S MATE RATING

This training manual is designed to aid the AG2 in preparing for advancement to AG1 and the AG1 in preparing for advancement to AGC. It is based primarily on the professional requirements or qualifications for AG1 and AGC, as specified in the Manual for Qualifications for Advancement, NAVPERS 18068-C.

In preparing for the advancement examination, this manual should be studied in conjunction with the manual, Military Requirements for Petty Officers 1 & C, NavTra 10057-C.

The intent of this chapter is to provide information on the enlisted rating structure, the AG rating, requirements and procedures for advancement, and references that will help you in performing your duties as an Aerographer's Mate. This chapter also includes information on how to make the best use of Rate Training Manuals. It is therefore strongly recommended that you study this chapter carefully before beginning intensive study of the remainder of the manual.

ENLISTED RATING STRUCTURE

The present enlisted rating structure includes two types of ratings: general ratings and service ratings.

GENERAL RATINGS are designed to provide paths of advancement and career development. A general rating identifies a broad occupational field of related duties and functions requiring similar aptitudes and qualifications. General ratings provide the primary means used to identify billet requirements and personnel qualifications. Some general ratings include service ratings; others do not. Both Regular Navy and Naval Reserve personnel may hold general ratings.

Subdivisions of certain general ratings are identified as SERVICE RATINGS. These service ratings identify areas of specialization within the scope of a general rating. Service ratings are established in those general ratings in which specialization is essential for efficient utilization of personnel. Although service ratings can exist at any petty officer level, they are most common at the PO3 and PO2 levels. Both Regular Navy and Naval Reserve personnel may hold service ratings

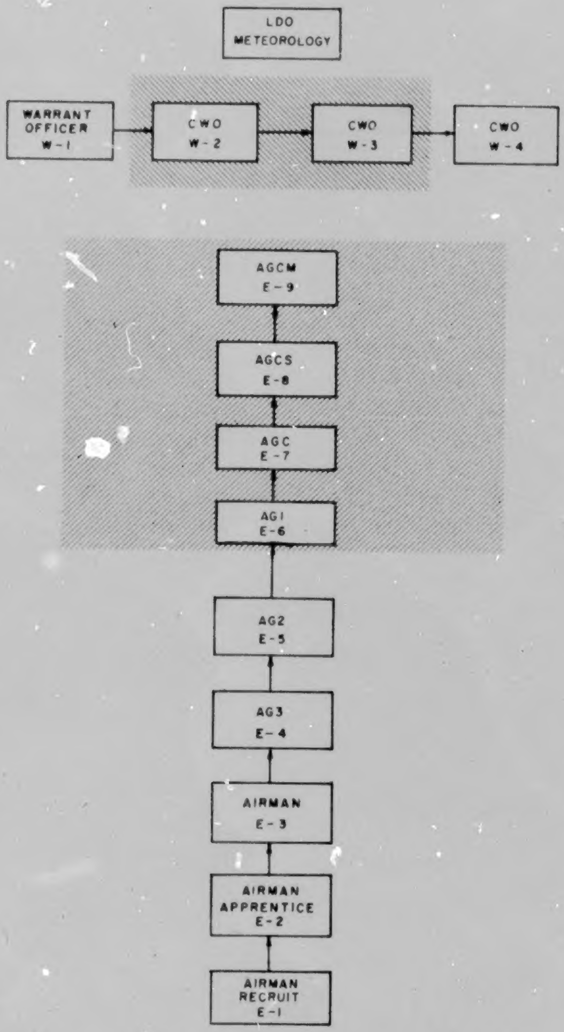
AEROGRAPHER'S MATE RATING

The Aerographer's Mate rating is divided into six rates or pay grades. It consists of just one rating—the general rating. The general rate is the pay grade level within the general rating. There is no service rating provided for Aerographer's Mates.

Figure 1-1 illustrates all paths of advancement for an Airman Recruit (AR) to Master Chief Aerographer's Mate (AGCM), or to Limited Duty Officer (LDO). Shaded areas indicate career stages where qualified enlisted men may advance to Warrant Officer (W-1), and selected Warrant Officers may advance to Limited Duty Officer.

Through the AG2 level, Aerographer's Mates are generally utilized in a variety of billets but are for the most part considered to be observer or plotters. With the advancement to the AG1 level their duties will change to the extent that they will act in the capacity of section leaders or assistants to the forecaster. It is also at this level that personnel, under supervision, will produce extended forecasts and other related products.

With advancement to AGC, personnel are expected to be proficient forecasters, a job they



AG.1

Figure 1-1.—Paths of advancement.

will continue to perform. They will also gain increased responsibility in the supervising of a larger number of personnel, or as in the case aboard ship be the division chief petty officer.

BILLET ASSIGNMENT

A wide variety of assignments, in addition to the normal shore and sea duty billets, becomes available at the AG1 and AGC level.

It should be pointed out that the successful completion of Aerographer's Mate, Class B school is a requirement for advancement to AGC and for authorization to be a meteorological/oceanographic forecaster. Forecaster qualifications are covered in Naval Weather Service Command Instruction 3140.5.

The AG School located at the Naval Air Technical Training Center, Lakehurst, N.J. has a large number of instructor billets at both the AG1 and AGC levels. Instructor billets are in the "A", "B", and "C" schools. Billets for support personnel such as in the testing unit are also available. Instructor billets are normally assigned on a voluntary basis and may be requested by qualified personnel via normal procedures.

A small number of AGC's may be assigned as Chief Petty Officer-in-Charge (CPO-IN-C) at smaller NWSED's.

Interesting and rewarding billets for personnel are found at the National Climatic Center, Asheville, N.C.; as aerial ice observers; and in a number of independent duty assignments on various staffs, ships, or stations.

It is impossible to list all the variety of duties that AG1's and AGC's can expect or have the opportunity of filling. It is anticipated, however, that with the intended shift to an all volunteer service increased authority and responsibility will be placed on senior enlisted personnel. It will become necessary to place even more emphasis on the selection of highly qualified personnel for advancement, and billet assignment.

The Aerographer's Mate detailer, assigned to the Bureau of Naval Personnel, through the use of official newsletters, notices, personal contact, and the command master chief petty officer, strives to keep personnel informed of changes that are anticipated (or that occur unexpectedly), which will have an effect on personnel assignment. In order for your duty preference to be known to your detailer, it is essential that your Enlisted Duty Preference form (NavPers 1306/63) be properly filled out, coded and submitted. In this manner personal considerations may be made by the detailer in filling the various billets available with qualified personnel.

NAVY ENLISTED CLASSIFICATION

The Navy Enlisted Classification (NEC) codes are utilized to a small extent within the Aerographer's Mate rating.

The NEC 7414 is used to indicate the holder has completed the course of instruction of the Rawin/Radiosonde Set Operators School, Class C. This NEC can also be obtained through on-the-job training.

The most important NEC at the AG2 and AG1 level is 7412, which indicates the holder is a graduate of Aerographer's Mate Class B school. Holders of this code can be qualified as meteorological/oceanographic forecasters within their commands.

Navy Enlisted Classification codes are also used to assign personnel to billets that require the special skills they have obtained.

RESPONSIBILITY

At the AG1 and AGC level personnel become acutely aware of the importance that weather plays in naval operations. It will be necessary to provide or assist in providing climatological statistics to personnel or staffs tasked with preparing operational plans and orders. As time for the commencement of operations draws near, the actual forecasts of expected conditions will be heavily relied upon for dictating the manner in which the operations will be conducted.

It is at this level, AG1 and AGC, that individuals begin to feel how responsible a position they have. They now have a more complete viewing of the overall picture. More complex questions than "What's the temperature out?" are directed toward them. They are accepted as the experts in their field.

LEADERSHIP

Since you have been a petty officer for some time, you realize that more leadership is required of the higher rates. Not only are you required to have superior knowledge, but you are also required to have the ability to handle personnel. This ability increases in importance as you advance through the various rates as a petty officer.

In General Order No. 21, the Secretary of the Navy outlined some of the most important aspects of naval leadership. Naval leadership means the art of accomplishing the Navy's mission through people. It is the sum of those qualities of intellect, of human understanding, and of moral character that enable a man to inspire and to manage a group of people successfully. Effective leadership, therefore, is based on personal example, good management practices, and moral responsibility. The term leadership includes all three of these elements.

The current Navy Leadership Program is designed to keep the spirit of General Order No. 21 ever before you. If the threefold objective is carried out effectively in every command, the program will make of you a better leader of men in your present billet and in your future assignments. As you advance up the leadership ladder, more and more your worth to the Navy will be judged on the basis of the amount of efficient work you obtain from your subordinates rather than how much of the actual work you do yourself.

For information on the practical application of leadership and supervision, study Military Requirements for Petty Officer 1 & C, NavTra 10057-C.

As you study this material containing leadership traits, keep in mind that probably none of our most successful leaders possessed all of these traits to a maximum degree, but a weakness in some traits was more than compensated for by strength in others. Critical self-evaluation will enable you to realize the traits in which you are strong, and to capitalize on them. At the same time you must constantly strive to improve on the traits in which you are weak.

Your success as a leader will be decided, for the most part, by your achievements in inspiring others to learn and perform. This is best accomplished by personal example.

ADVANCEMENT

By this time, you are probably well aware of the personal advantages of advancement—higher pay, greater prestige, more interesting and challenging work, and the satisfaction of getting ahead in your chosen career. By this time, also, you have probably discovered that one of the most enduring rewards of advancement is the training you acquire in the process of preparing for advancement.

The Navy also profits by your advancement. Highly trained personnel are essential to the functioning of the Navy. By advancement, you increase your value to the Navy in two ways: First, you become more valuable as a person who can supervise, lead, and train others and second you become more valuable as a technical specialist and thus make far-reaching contributions to the entire Navy.

Since you are studying for advancement to PO1 or CPO, you are probably already familiar with the requirements and procedures for advancement. However, you may find it helpful to read the following sections. The Navy does not stand still. Things change all the time, and it is possible that some of the requirements have changed since the last time you went up for advancement. Furthermore, you will be responsible for training others for advancement; therefore, you will need to know the requirements in some detail.

HOW TO QUALIFY FOR ADVANCEMENT

To qualify for advancement, a person must:

1. Have a certain amount of time in grade.
2. Complete the required military and professional training manuals.
3. Demonstrate the ability to perform all the PRACTICAL requirements for advancement by completing applicable portions of the Record of Practical Factors, NavEdTra 1414/1 AG.
4. Be recommended by his commanding officer.
5. Demonstrate his KNOWLEDGE by passing a written examination on (a) military requirements, and (b) professional qualifications.

Remember that the requirements for advancement can change. Check with your educational services office to be sure that you know the most recent requirements.

When you are training lower rated personnel, it is a good idea to point out that advancement is not automatic. Meeting all the requirements makes a person ELIGIBLE for advancement, but it does not guarantee his advancement. Such factors as the score made on the written examination, length of time in service, performance marks, and quotas enter into the final determination of who will actually be advanced.

HOW TO PREPARE FOR ADVANCEMENT

Preparation for advancement includes studying the qualifications, working on the practical factors, studying the required Rate Training Manuals, and studying any other material that may be specified. To prepare yourself for advancement or to help others prepare for advancement, you will need to be familiar with (1) the "Quals" Manual, (2) the Record of Practical Factors, NavEdTra 1414/1, (3) a NavEdTra publication called Bibliography for Advancement Study NavEdTra 10052 (Series), and (4) Rate Training Manuals. The following sections describe these materials and give some information on how to use them to the best advantage.

"Quals" Manual

The Manual of Qualifications for Advancement, NavPers 18068 (Series), gives the minimum requirements for advancement to each rate within each rating. This manual is usually called the "Quals" Manual, and the qualifications themselves are often called "quals." The qualifications are of two general types: (1) military requirements, and (2) professional or technical qualifications. Military requirements apply to all ratings rather than to any one rating alone. Professional qualifications are technical or professional requirements that are directly related to the work of each rating.

Both the military requirements and the professional qualifications are divided into subject matter groups. Then, within each subject matter group, they are divided into PRACTICAL FACTORS and KNOWLEDGE FACTORS.

The qualifications for advancement and a bibliography of study materials are available in your educational services office. The "Quals" Manual is changed more frequently than Rate Training Manuals are revised. By the time you are studying this training manual, the "quals" may have been changed. Never trust any set of "quals" until you have checked the change number against an UP-TO-DATE copy of the "Quals" Manual.

In training others for advancement, emphasize these three points about the "quals":

1. The "quals" are the MINIMUM requirements for advancement. Personnel who study MORE than the required minimum will have a great advantage when they take the written examinations for advancement.

2. Each "qual" has a designated rate level—chief, first class, second class, or third class. You are responsible for meeting all "quals" specified for the rate level to which you are seeking advancement AND all "quals" specified for lower rate levels.

3. The written examinations for advancement will contain questions relating to the practical factors AND to the knowledge factors of BOTH the military requirements and the professional qualifications.

Record of Practical Factors

A special form known as the Record of Practical Factors, NavEdTra 1414/1, is used to record the satisfactory performance of the practical factors. This form lists all military and all professional practical factors. Whenever a person demonstrates his ability to perform a practical factor, appropriate entries must be made in the DATE and INITIAL columns. As a PO1 or CPO, you will often be required to check the practical factor performance of lower rated personnel and to report the results to your supervising officer.

As changes are made periodically to the "Quals" Manual, new forms of NavEdTra 1414/1 are provided when necessary. Extra space is allowed on the Record of Practical Factors for entering additional practical factors as they are published in changes to the "Quals" Manual. The Record of Practical Factors also provides space for recording demonstrated proficiency in skills which are within the general scope of the rate but which are not identified as minimum qualifications for advancement. Keep this in mind when you are training and supervising other personnel. If a person demonstrates proficiency in some skill which is not listed in the "quals" but which is within the general scope of the rate, report this fact to the supervising officer so that an appropriate entry can be made in the Record of Practical Factors.

When you are transferred, the Record of Practical Factors should be forwarded with your service record to your next duty station. It is a good idea to check and be sure that this form is actually inserted in your service record before you are transferred. If the form is not in your record, you may be required to start all over again and requalify in practical factors that have already been checked off. You should also take some responsibility for helping lower rated personnel keep track of their practical factors records when they are transferred.

NavEdTra 10052

Bibliography for Advancement Study, NavEdTra 10052 (Series) is a very important publication for anyone preparing for advancement. This publication lists required and recommended Rate Training Manuals and other reference material to be used by personnel working for advancement. NavEdTra 10052 (Series) is revised and issued once each year by the Chief of Naval Education and Training Support. Each revised edition is identified by a letter following the NavEdTra number. When using this publication, be SURE you have the most recent edition.

The required and recommended references are listed by rate level in NavEdTra 10052 (Series). It is important to remember that you are responsible for all references at lower rate levels, as well as those listed for the rate to which you are seeking advancement.

Rate Training Manuals that are marked with an asterisk (*) in NavEdTra 10052 (Series) are MANDATORY at the indicated rate levels. A mandatory training manual may be completed by (1) passing the appropriate Non-Resident Career Course that is based on the mandatory training manual; (2) passing locally prepared tests based on the information given in the mandatory training manual; or (3) in some cases, successfully completing an appropriate Navy school.

When training personnel for advancement, do not overlook the section of NavEdTra 10052 (Series) which lists the required and recommended references relating to the military requirements for advancement. All personnel must complete the mandatory military requirements training manual for the appropriate rate level before they can be eligible to advance. Also, make sure that personnel working for advancement study the references which are listed as recommended but not mandatory in NavEdTra 10052 (Series) may be used as source material for the written examinations, at the appropriate levels.

Rate Training Manuals

There are two general types of Rate Training Manuals. Manuals (such as this one) are prepared for most enlisted rates and ratings, giving information that is directly related to the professional qualifications for advancement. Basic manuals give information that applies to more than one rate and rating.

Rate Training Manuals are revised from time to time to bring them up to date. The revision of a Rate Training Manual is identified by a letter following the NavEdTra number. You can tell whether a Rate Training Manual is the latest edition by checking the NavEdTra number (and the letter following the number) in the most recent edition of List of Training Manuals and Correspondence Courses, NavEdTra 10061 (Series).

Rate Training Manuals are designed for the special purpose of helping naval personnel prepare for advancement. By this time, you have probably developed your own way of studying these manuals. Some of the personnel you train, however, may need guidance in the use of Rate Training Manuals. Although there is no single "best" way to study a training manual, the following suggestions have proved useful for many people.

1. Study the military requirements and the professional qualifications for your rate before you study the training manual, and refer to the "quals" frequently as you study. Remember, you are studying the training manual primarily to meet these "quals."

2. Before you begin to study any part of the training manual intensively, get acquainted with the entire manual. Read the preface and the table of contents. Check through the index. Thumb through the manual without any particular plan, looking at the illustrations and reading bits here and there as you see things that interest you.

3. Look at the training manual in more detail, to see how it is organized. Look at the table of contents again. Then, chapter by chapter, read the introduction, the headings, and the subheadings. This will give you a pretty clear picture of the scope and content of the manual.

4. When you have a general idea of what is in the training manual and how it is organized, fill in the details by intensive study. In each study period, try to cover a complete unit—it may be a chapter, a section of a chapter, or a subsection. The amount of material you can cover at one time will vary. If you know the subject well, or if the material is easy, you can cover quite a lot at one time. Difficult or unfamiliar material will require more study time.

5. In studying each unit, write down questions as they occur to you. Many people find it helpful to make a written outline of the unit as they study, or at least to write down the most important ideas.

6. As you study, relate the information in the training manual to the knowledge you already have. When you read about a process, a skill, or a situation, ask yourself some questions. Does this information tie in with past experience? Or is this something new and different? How does this information relate to the qualifications for advancement?

7. When you have finished studying a unit, take time out to see what you have learned. Look back over your notes and questions. Without looking at the training manual, write down the main ideas you have learned from studying this unit. Do not just quote the book. If you cannot give these ideas in your own words, the chances are that you have not really mastered the information.

8. Use Non-Resident Career Courses whenever you can. The non-resident career courses are based on the Rate Training Manuals or other appropriate texts. As mentioned before, completion of a mandatory Rate Training Manual can be accomplished by passing a Non-Resident Career Course based on the training manual. You will probably find it helpful to take other non-resident career courses, as well as those based on mandatory training manuals. Taking a

non-resident career course helps you master the information given in the training manual, and also gives you an idea of how much you have learned.

SOURCES OF INFORMATION

As a PO1 or CPO, you must have an extensive knowledge of the references to consult for accurate, authoritative, up-to-date information on all subjects related to the military and professional requirements for advancement.

Publications mentioned in this chapter are subject to change or revision from time to time—some at regular intervals, others as the need arises. When using any publication that is subject to revision, make sure that you have the latest edition. When using any publication that is kept current by means of changes, be sure you have a copy in which all official changes have been made.

A list of training manuals and publications that will be helpful as references and for additional study in preparing for advancement is included in the reading list at the beginning of this text. Additional training manuals that are applicable are available through your educational services officer.

In addition to training manuals and publications, training films furnish a valuable source of supplementary information. Films that may be helpful are listed in the U.S. Navy Film Catalog, NavAir 10-1-777.

CHAPTER 2

WORLD CLIMATE AND WEATHER

When a forecaster limits himself to shortrange considerations of the weather he is at the same time limiting his forecast. He must not only consider the present synoptic indications, but must carefully study the large-scale climatic features and seasonal climatic changes as well. The relationships between the current meteorological structure and the climatic background of the area are of great importance to the critical forecaster.

This chapter will give you an introduction to climate, and the elements which are used to assemble the data obtained and to classify this data into climatic zones and types. Many other excellent texts and references are available for a further study of the characteristics of climates and their importance in weather forecasting.

CLIMATE AND CLIMATOLOGY

CLIMATE

Climate is defined as the average or collective state of the earth's atmosphere at any given location or area within a specified period of time. We think of weather as the day-to-day changes in the atmosphere. On the other hand, the climate of an area is determined over periods of many years and represents the general weather characteristics of an area or locality.

CLIMATOLOGY

Climatology is the scientific study of climate. It is a branch of meteorology, or simply the study of the atmosphere. There are three principal approaches to the study of climatology. They are physical, descriptive, and dynamic.

Physical Climatology

This approach to climatology seeks to explain the causes of the differences in climate in the light of the physical processes influencing climate, and the processes producing the various kinds of physical climates such as marine, deserts, mountain, etc.

Descriptive Climatology

Descriptive climatology typically orients itself in terms of geographic regions and is also referred to as regional climatology. A description of the various types of climates is made on the basis of analyzed statistics from a particular area. A further attempt is made to describe the interaction of weather and climatic elements upon the people and the areas under consideration.

Dynamic Climatology

This study of climate attempts to relate characteristics of the general circulation of the atmosphere to climate.

Climatology as Related to Other Sciences

One of three prefixes is often added to the word "climatology" to denote a scale or magnitude. Micro, meso, and macro indicate small, medium, and large scales, respectively.

Microclimatological studies often measure contrasts between hilltop and valley, city and surrounding country, or they may be of an extremely small scale—one side of a hedge contrasted with the other, a plowed furrow versus level soil, or opposite leaf surfaces. Climate in the microscale may be effectively modified by relatively simple human efforts.

Macroclimatology is the study of the large-scale climate of a large area or country. Climate of this type is not so easily modified by small human efforts.

Mesoclimatology embraces a rather indistinct middle ground between macroclimatology and microclimatology. The areas are smaller than those of macroclimatology and larger than those of microclimatology and may or may not be climatically representative of a general region.

These terms (micro, meso, and macro) are also applied to meteorology.

Climate has become increasingly important in other scientific fields. Geographers, hydrologists, and oceanographers use quantitative measures of climate to describe or analyze the influence of our atmospheric environment. Climate classification has developed primarily in the field of geography. The basic role of the atmosphere in the "hydrologic cycle" is an essential part of the study of hydrology. Parallel to this, both air and water measurements are required to understand the energy exchange between air and ocean.

CLIMATIC ELEMENTS

The weather elements which are used to describe climate are discussed in the following section. A further discussion of the effects upon some of these elements are covered later in this chapter.

TEMPERATURE

This element is undoubtedly the most important of all the climatic elements. The temperature of an area or locality is dependent upon latitude, or the distribution of incoming and outgoing radiation; nature of the surface (land or water); altitude; and the prevailing winds. The temperature normally used in climatology is the surface temperature.

HYDROMETEORS (PRECIPITATION)

This is the second most important climatic element. In most studies, this includes all water reaching the earth's surface by falling either in liquid or in solid state. The most significant forms are rain, snow, and hail. Precipitation has a wide range of variability over the surface of the earth, and because of this variability a longer series of observations is generally required to establish a mean or an average. Since two stations could have the same amount of annual precipitation, but it could occur in different months, or days during these months, and the intensity could also vary, it often becomes necessary to include such factors as average number of days with precipitation, average amount per day and other factors.

Further, since precipitation amounts are directly associated with amount and type of clouds, cloud cover must also be included along with a precipitation study. Cloud cover is usually expressed in tenths of sky cover. Precipitation is expressed in most studies in inches, but centimeters may be used in some studies which are based on data derived from the metric system. Cloud climatology also includes such phenomena as fog and thunderstorms.

WIND

Climatologists are mostly interested in wind in terms of wind direction, speed, and gustiness. Frequently it is expressed in terms of "prevailing" wind direction, average speeds, and maximum gusts. Some climatological studies use "resultant" wind which is the vectorial average of all wind directions and speeds for a given level, at a specific place, and for a given period. The vectorial average is obtained by dividing each wind observation into components, making a summation for a given period, then obtaining averages and converting the average components into a single vector.

CONDENSATION

This climatic element includes such deposits as dew, frost, and rime ice. It is not of particular importance in most general studies where there is sufficient rain to support life. In some areas, however, it can be important.

EVAPORATION

Although this climatic element does not receive the attention it deserves, it can be extremely important when it is considered in relation to the formation of weather phenomena over water bodies or oceanic areas. It can be an important factor in the formation of fogs over such areas.

EXPRESSION OF CLIMATIC ELEMENTS

As an Aerographer's Mate, you must understand climatological terms and the methods used to derive these terms, in order for the climatological data to have definite meaning to you. In this section the most commonly used terms are defined, and where applicable, their usage in expressing climatic elements is explained.

MEAN OR AVERAGE

The mean is the most commonly used climatological parameter. The term "mean" normally refers to the arithmetic mean which is obtained in the same manner as the average. This is an average obtained by adding the values of all factors or cases and then dividing by the number of items. For example, the average daily temperature would be the sum of the hourly temperatures divided by 24. The mean, as computed in this manner, is generally optimum for both the expected value and the center of the distribution for temperature.

Other methods are used for computing various meteorological elements. For example, the mean temperature for the day has been derived by simply adding the maximum and minimum values for the day and dividing by 2. Assume the maximum temperature for a certain day is 75°F and the minimum temperature is 57°F; the mean temperature for the day is 66°F.

Unfortunately the term "mean" has been used in many climatological records without clarification as to how it was computed. In most cases, the difference in results obtained is slight. In analyzing weather data, the terms "average" and "mean" are often used interchangeably.

NORMAL

In climatology, the term "normal" is applied to the average value which in the course of a period of time any meteorological element is found to have on a specified date or during specified times. These times may be a particular month or other portion of the year. They may refer to a season or to a year as a whole. The normal serves as a standard with which values occurring on a date or during a specified time may be compared.

ABSOLUTE

The term "absolute" usually is applied in climatology to the extreme highest and lowest values for any given meteorological element which has been recorded at the place of observation. Assume, for example, that the extreme highest temperature ever recorded at a particular station was 106°F and the lowest recorded was -15°F. These are called the absolute maximum and absolute minimum, respectively.

EXTREMES

The term "extreme" is applied to the highest value and the lowest value for a particular meteorological element which have occurred over a period of time. The term is usually applied to months, seasons, years, or a number of years. The term may be used for a calendar day only, for which it is particularly applicable to temperature. For example, the highest and lowest temperature readings for a particular day are considered the temperature extremes for that day. At times it is applied to the average of the highest and lowest temperatures and termed mean monthly extremes and mean annual extremes.

RANGE

Range is the difference between the highest and lowest values and reflects the extreme variations of these values. This statistic is not recommended except for very crude work, since it has a high variability. The range is related to the extreme values of record and can be useful in determining the extreme range for the records available.

FREQUENCY

Frequency is defined as the number of times a certain value occurs within a specified period of time. When a large number of variate values need to be presented, a condensed presentation of data may be obtained by means of a frequency distribution.

MODE

The mode is defined as the value which occurs with the greatest frequency, or the value about which the most cases occur.

The mode cannot be determined readily from unorganized data; therefore, the data must be grouped in a frequency distribution before its location can be determined accurately. It is not always well defined or possible to locate properly. The maximum density point may be more than one point so that the point determined as the mode depends upon the judgment or desire of the person interpreting and using the mode.

In general, the mode is not recommended for climatology. However, it can be useful in local climatological studies or in determining the most common or frequently attained value for a prediction technique. For example, in the objective technique, The Prediction of Maritime Cyclones, the deepening prediction graph for cyclones, has a modal value as well as a maximum value.

MEDIAN

The median is the value at the midpoint in an array. For determining the median all items have to be arranged in order of size. Rough estimates of the median may be obtained by taking the middle value of an ordered series or if there are two middle values, they may be averaged to obtain the median. The position of the median may be found by the use of the following formula:

$$\text{Median} = \frac{n + 1}{2}$$

where n is the number of items.

The median is not widely used in climatological computations. Some writers recommend the use of the median instead of the mean or average for some of the climatic elements because the extremes of these elements are averaged and combine in some cases to give a wholly unrepresentative picture of distribution and probability of that particular element. However, a longer period of record might be required to formulate a median.

STANDARD DEVIATIONS

In many analyses of climatological data, it is desirable to compute deviation of all items from a central point. This may be obtained from a computation of either the mean (or average) deviation or the standard deviation. These are termed measures of dispersion and are used to determine whether the average is truly representative or to determine the extent by which data vary from the average.

The average deviation is obtained by computing the arithmetic average of the deviations from an average of the data. First, we obtain an average of the data, then the deviations of the individual items from this average are determined, and finally the arithmetic average of these deviations is computed. The plus or minus signs are disregraded. The formula for computation of the average deviation is as follows:

$$\text{Average Deviation} = \frac{\Sigma d}{n}$$

the Greek letter Σ (sigma) means the sum of "d" which are the deviations and "n" is the number of items.

The standard deviation, like the average deviation, is the measure of the scatter or spread of all values in a series of observations. To obtain the standard deviation, each deviation from the arithmetic average of the data is squared. Next determine the arithmetic average of the squared deviations, finally extracting the square root of this average. This is also called the root mean square deviation, in that it is the square root of the mean of the deviations squared.

The formula for computing standard deviation is given as follows:

$$\text{Standard Deviation} = \sqrt{\frac{\Sigma d^2}{n}}$$

Σd^2 is the sum of the squared deviations from the arithmetic average, and "n" is the number of items in the group of data.

An example of the computation of both the average deviation and the standard deviation is given in table 2-1 and in the following paragraphs.

Suppose, on the basis of 10 years of data (1954-1963), for the month of January we wished to compute the average deviation of mean temperature and the standard deviation for this period. First arrange the data in tabular form (such as in table 2-1), giving the year in the first column, the mean monthly temperature in the second column, the deviations from an arithmetic average of the mean temperature in the third column, and the deviations from the mean (column 3) squared in column 4.

Table 2-1.—Computation of average and standard deviation.

January year	Mean temperature	Deviations from mean	Deviations squared
1954	47	-4	16
1955	51	+0	0
1956	53	+2	4
1957	50	-1	1
1958	49	-2	4
1959	55	+4	16
1960	46	-5	25
1961	52	+1	1
1962	57	+6	36
1963	50	-1	1
Totals	510	26	104
Mean	51	2.6	3.2

To compute the average deviation:

1. Add all the temperatures in column 2 and divide by the number of years (10 in this case) to get the arithmetic average of temperature.

2. In column 3, compute the deviation from the mean or average determined in step 1. (The mean temperature for the 10-year period was 51°F.)

3. Total column 3 (disregarding the negative and positive signs). This total is 26.

4. Apply the formula for Average Deviation,

$$\frac{\Sigma d}{n} = \frac{26}{10} = 2.6°\text{F}$$

The average deviation of temperature during the month of January for the period of record, 10 years, is 2.6°F.

To compute the standard deviation:

1. Square the deviations from the mean (column 3).

2. Total these squared deviations. In this case the total is 104.

3. Apply the formula for standard deviation:

$$\text{Standard Deviation} = \sqrt{\frac{\Sigma d^2}{n}} = \sqrt{\frac{104}{10}}$$

$$\sqrt{10.4} = 3.225 \text{ or } 3.2°\text{F}$$

Thus the standard deviation of temperature for the month and period in question is 3.2°F (rounded off to the nearest one tenth degree).

Naturally a question arises. Just what use is this and how can I apply the computations made in table 2-1 for everyday forecasting purposes? With the standard deviation just determined it is readily apparent that the implication is that there is a small range of mean temperature during this month. If we had available a frequency distribution of temperature for this station for each day of the month, we could readily determine the percentage of readings which would fall in the 6.4-degree spread (3.2 either side of the mean). From these data we could then formulate a probability forecast or the number of days within this range that we could expect the normal or mean temperature to occur. This study could further be broken down into hours of the day, etc.

CLASSIFICATION OF CLIMATE

The climate of a given region or locality is determined by a combination of several meteorological elements, and not just one element alone. For example, two regions may have similar temperature climates but very different precipitation climates. Their climatic difference therefore becomes apparent only if more than one climatic factor is considered.

Since the climate of a region is composed of all the averages of the various climatic elements, such as dew, ice, rain, temperature, wind force, and wind direction, it is obvious that no two locations have exactly the same climate. However, it is possible to place similar areas into a grouping known as a climatic zone.

CLIMATIC ZONES

The basic grouping of climatic zones consists of classifying climates into five broad belts based on astronomical or mathematical grounds. Actually they are zones of sunshine, or solar climate. The five basic regions or zones are the Torrid or Tropical Zone, the two Temperate Zones, and the two Polar Zones. The Tropical Zone is limited on the north by the Tropic of Cancer and on the south by the Tropic of Capricorn, which are located at 23 1/2° N and S lat, respectively. The Temperate Zone of the Northern Hemisphere is limited on the south by the Tropic of Cancer and on the north by the Arctic Circle, which is located at 66 1/2° N lat. The Temperate Zone of the Southern Hemishpere is bounded on the north by the Tropic of Capricorn and on the sourth by the Antarctic Circle, which is located at 66 1/2° S lat. The two Polar Zones are the areas in the polar regions which have the Arctic and Antarctic Circles as their boundaries. The Polar Zones are sometimes called the Frigid Zones.

A glance at any chart depicting the isotherms over the surface of the earth will show that the isotherms do not coincide with latitude lines. In fact, at some places the isotherms parallel the longitude lines more closely than they parallel the latitude lines. The astronomical or light zones therefore differ from the zones of heat. A closer approach to the understanding of climate can be made if the climatic zones are limited by isotherms rather than by parallels of latitude. (See fig. 2-1.)

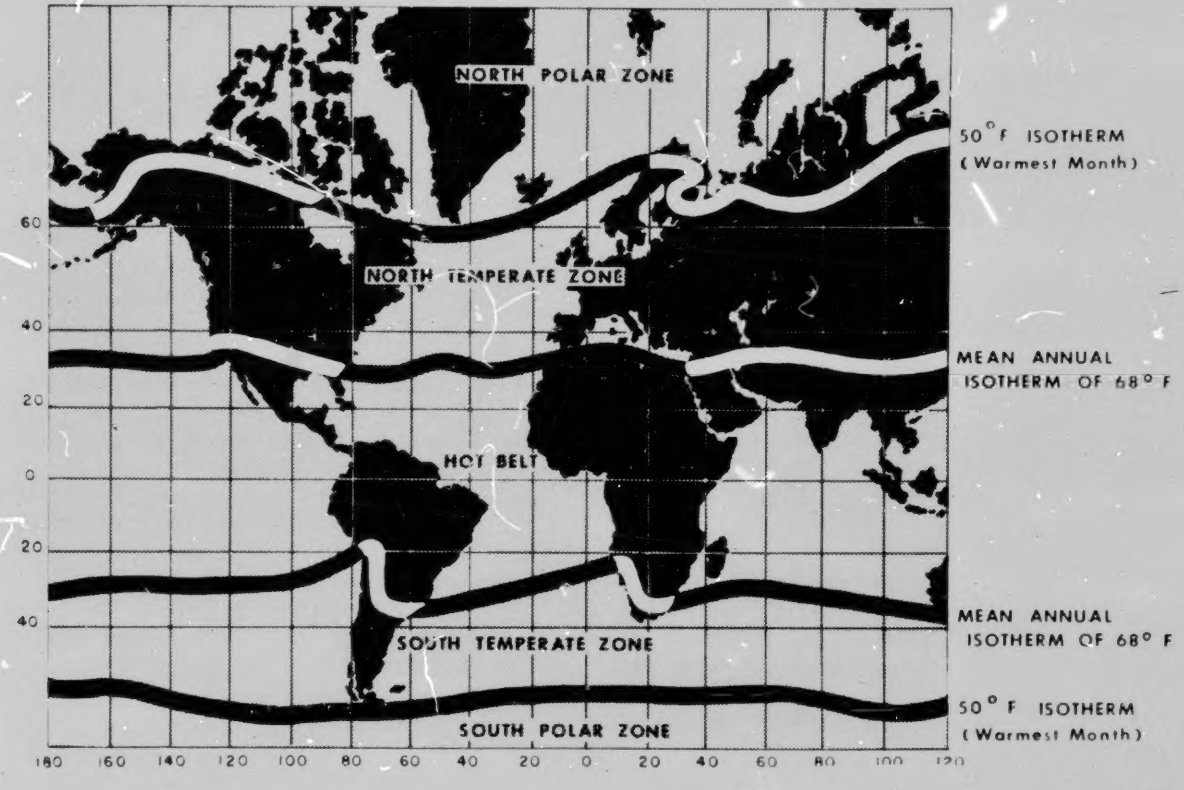

NORTH POLAR ZONE
50°F ISOTHERM
(Warmest Month)
60
NORTH TEMPERATE ZONE
40
MEAN ANNUAL
ISOTHERM OF 68° F
20
HOT BELT
0
20
40
MEAN ANNUAL
ISOTHERM OF 68° F
SOUTH TEMPERATE ZONE
50° F ISOTHERM
(Warmest Month)
SOUTH POLAR ZONE
180
160
140
120
100
80
60
40
20
0
20
40
60
80
100
120

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Figure 2-1.—Temperature zones.

CLIMATIC TYPES

Any classification of climate depends to a large extent on the purpose of the classification. A classification for the purpose of establishing air stations, for instance, where favorable flying conditions are important, would differ considerably from one for establishing the limits of areas that are favorable for the growing of crops. There are two classifications that particularly merit your attention. They are the classifications of Koeppen and Thornthwaite.

In Koeppen's classification there are five main climatic types. They are TROPICAL RAIN, DRY, HUMID MESOTHERMAL, HUMID MICROTHERMAL, and POLAR. These main types are further divided into climatic provinces. The Koeppen classification is based mainly on temperature, precipitation amount, and season of maximum precipitation. Numerical values of these elements constitute the boundaries of the above types and were selected primarily according to their effect on plant growth.

Thornthwaite's classification of climates places a great deal of emphasis on the effectiveness of precipitation. Effectiveness of precipitation means the relationship between precipitation and evaporation at a certain locality. Thornthwaite classified climates into five climatic provinces, which are WET, HUMID, SUBHUMID, SEMIARID, and ARID. To each of the provinces is given a precipitation effectiveness rating.

CLIMATIC CONTROLS

The variation of climatic elements from place to place and from season to season is caused by several factors called climatic controls. The same basic factors that cause weather in the atmosphere also determine the climate of an area. These controls, acting in different combinations and with varying intensities act upon temperature, precipitation, humidity, air pressure, and winds to produce many types of weather and therefore climate.

Four factors largely determine the climate of every ocean and continental region. They are as follows:

1. Latitude.
2. Land and water distribution.
3. Topography.
4. Ocean currents.

LATITUDE

Perhaps no other climatic control has such a marked effect upon climatic elements as does the latitude, or the position of the earth relative to the sun. The angle at which rays of sunlight reach the earth and the number of "sun" hours each day depend upon the distance from the Equator. (See fig. 2-2.) Therefore, the extent to which an air mass is heated is influenced by the latitude. Latitude influences the sources and direction of air masses and the weather they bring with them.

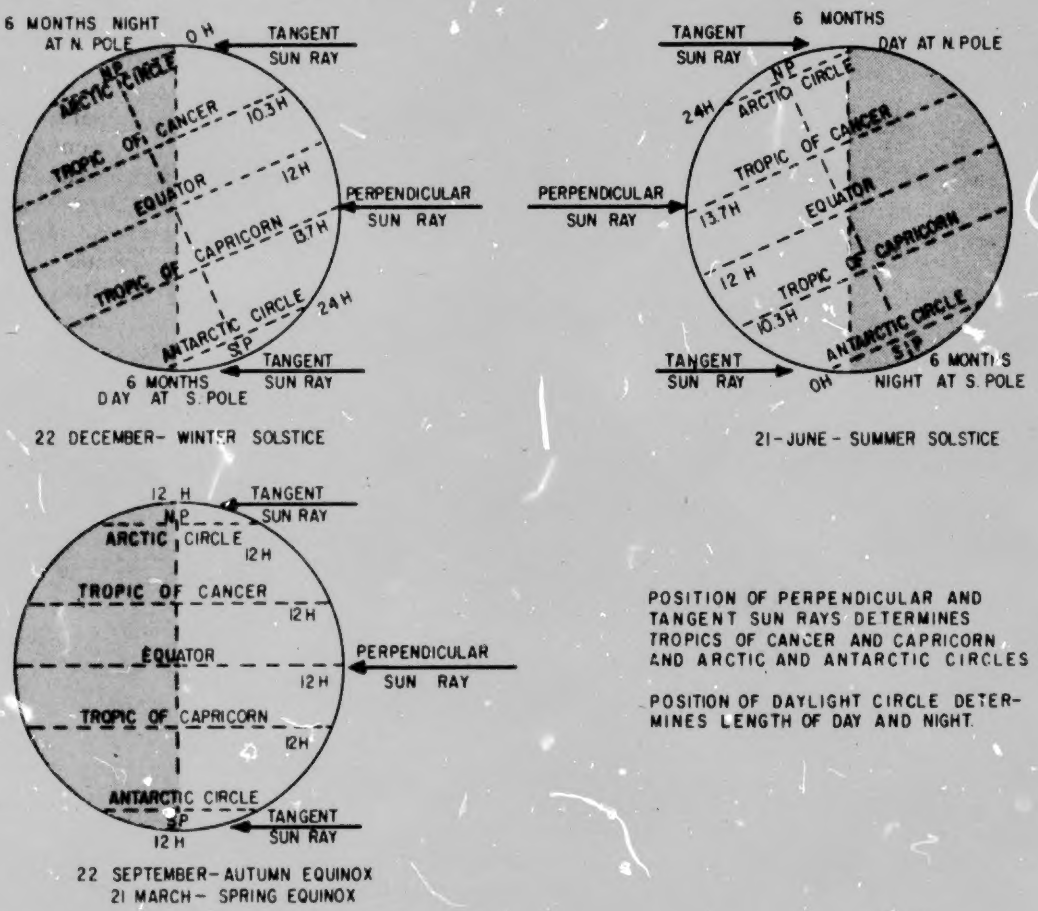
Influence on Air Temperature

Regions under direct or nearly direct rays of the sun receive more heat (per unit of time) than those under oblique rays. The heat brought about by the slanting rays of early morning may be compared with the heat that is caused by the slanting rays of winter. The heat which is due to the more nearly direct rays of midday may be compared with the heat resulting from the more nearly direct rays of summer.

The length of the day, like the angle of the sun's rays, influences the temperature. The length of the day varies with the latitude and the season of the year. A place near the Equator has about 12 hours of daylight every day in the year. Because of this, and because the sun at noonday is always high in the sky (giving nearly direct rays), equatorial regions do not have pronounced seasonal temperature changes.

During summer in the Northern Hemisphere all places north of the Equator have more than 12 hours of daylight. This situation is reversed in the winter; latitudes north of the Equator received less than 12 hours of daylight.

Great seasonal variation in the length of the day and the seasonal difference in the angle at which the sun's rays reach the earth's surface cause seasonal temperature differences in middle and high latitudes. In the far north, long hours of winter darkness produce cold temperatures that breed powerful polar air masses; conversely long hours of summer daylight weaken the polar air masses.


6 MONTHS NIGHT AT N. POLE
0 H
TANGENT SUN RAY
NP
ARCTIC CIRCLE
TROPIC OF CANCER
10.3 H
12 H
EQUATOR
PERPENDICULAR SUN RAY
TROPIC OF CAPRICORN
13.7 H
ANTARCTIC CIRCLE
24 H
SP
TANGENT SUN RAY
6 MONTHS DAY AT S. POLE
22 DECEMBER- WINTER SOLSTICE
TANGENT SUN RAY
6 MONTHS DAY AT N. POLE
NP
ARCTIC CIRCLE
24H
TROPIC OF CANCER
PERPENDICULAR SUN RAY
13.7H
EQUATOR
12 H
TROPIC OF CAPRICORN
10.3H
ANTARCTIC CIRCLE
SP
TANGENT SUN RAY
0H
6 MONTHS NIGHT AT S. POLE
21-JUNE - SUMMER SOLSTICE
12 H
TANGENT SUN RAY
NP
ARCTIC CIRCLE
12 H
TROPIC OF CANCER
12 H
EQUATOR
PERPENDICULAR SUN RAY
12 H
TROPIC OF CAPRICORN
12H
ANTARCTIC CIRCLE
SP
12 H
TANGENT SUN RAY
22 SEPTEMBER- AUTUMN EQUINOX
21 MARCH- SPRING EQUINOX

POSITION OF PERPENDICULAR AND TANGENT SUN RAYS DETERMINES TROPICS OF CANCER AND CAPRICORN AND ARCTIC AND ANTARCTIC CIRCLES

POSITION OF DAYLIGHT CIRCLE DETERMINES LENGTH OF DAY AND NIGHT.

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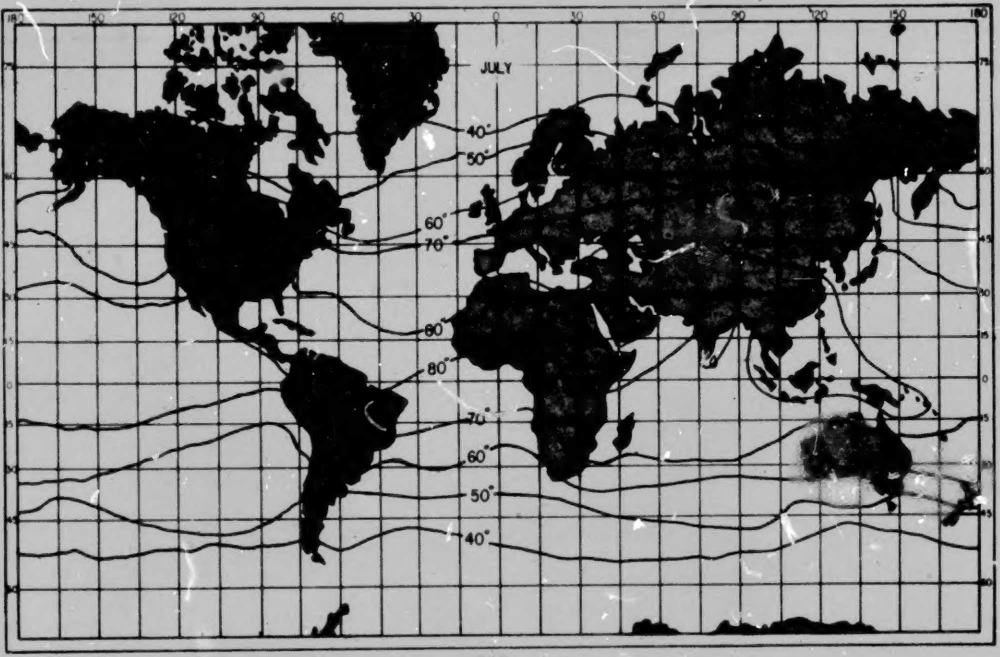
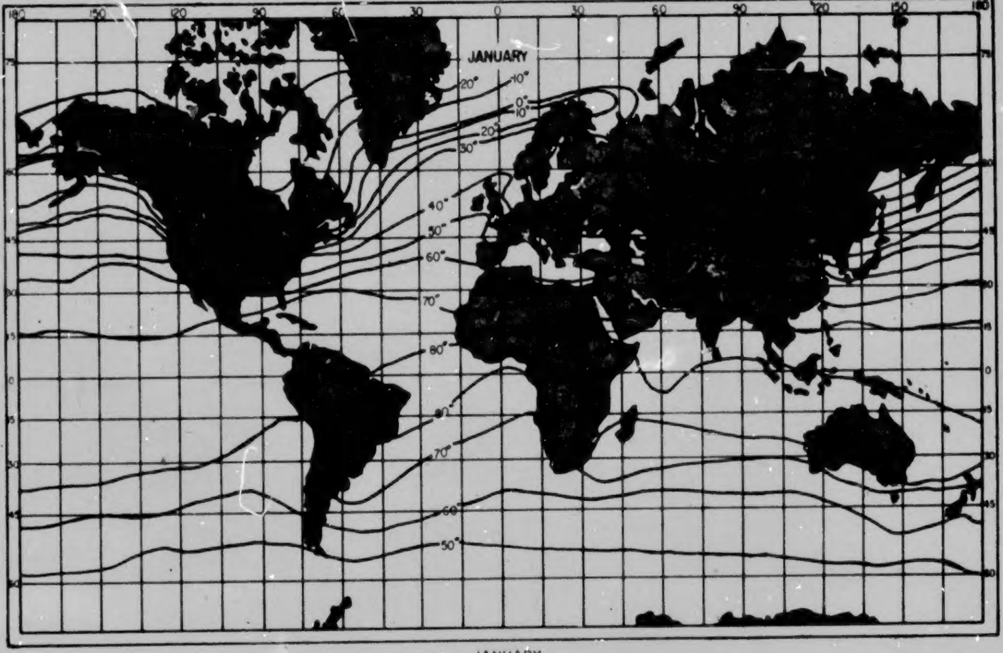
Figure 2-2.—Latitude differences in amount of insolation.

The hot and humid climates of equatorial Africa and South America are good examples of the influence that latitude has on climate. At no time during the year are the sun's rays at much of an oblique angle. Therefore, there is little difference between mean temperature for the coldest and warmest month. Contrast this picture with the opposite extreme, where the sun is either below the horizon for a great deal of the time or is only slightly above the horizon at any time. The sun's rays in reaching the earth's surface in polar regions make such a small angle with the earth's surface that the energy received per unit area is extremely small and the sun's effectiveness is minimized even though it may shine for days without ceasing.

The average world surface temperatures are represented on two world charts for July and January as shown in figure 2-3. It must be remembered that these are mean charts and are not meant to be an accurate portrayal of the temperatures on any one particular day. Note that in general the temperatures decrease from low to high latitudes. This is the latitude factor.


180
150
120
90
60
30
0
30
60
90
120
150
180
JULY
40°
50°
60°
70°
80°
80°
70°
60°
50°
40°

JULY


180
150
120
90
60
30
0
30
60
90
120
150
180
JANUARY
20°
10°
0°
10°
20°
30°
40°
50°
60°
70°
80°
80°
70°
60°
50°

JANUARY

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Figure 2-3.—World distribution of mean temperatures (degrees Fahrenheit).

LAND AND WATER DISTRIBUTION

Because land heats and cools about four times faster than water the location of continents and oceans greatly alters the earth pattern of air temperature and influences the sources and direction of movement of air masses.

Influence on Air Temperature

Coastal areas take on the temperature characteristics of the land or water to their windward. In latitudes of prevailing westerly winds, for example, west coasts of continents have oceanic temperatures and east coasts have continental temperatures. The temperatures are determined by the windflow.

Since the upper layers of the ocean are nearly always in a state of violent stirring, heat losses or heat gains occurring at the sea surface are distributed throughout a large volume of water. This mixing process sharply reduces the temperature contrasts between day and night and between winter and summer over oceanic areas.

Over land, there is no redistribution of heat by turbulence, and the effect of conduction is negligible. Thus violent contrasts between seasons and between day and night are created in the interiors of continents. During winter, a large part of the incident solar radiation is reflected back toward space by the snow cover that extends over large portions of the northern continents. For these reasons, the northern continents serve as manufacturing plants for dry polar air. The polar air cap is no longer symmetrical, but is displaced far to the south, particularly over the interior of Asia.

The large temperature difference between the land and water surface, which reverses between the two seasons, determines to a great extent the seasonal weather patterns.

You will note in figure 2-3 that in the Northern Hemisphere the isotherms are more closely spaced and parallel in winter. In the Southern Hemisphere, the temperature gradient does not have as great a seasonal change as it does in the Northern Hemisphere. This is due to the unequal distribution of land and water on the two hemispheres. Since the Southern Hemisphere has less land and more water surface than the Northern Hemisphere, the change due to the greater water surface is less with consequent more nearly uniform isotherms. Too, the continents of the Southern Hemisphere taper toward the poles and do not extend to as high a latitude as in the Northern Hemisphere.

The nature of the surface affects the local heat distribution. Color, texture, and vegetation influence the rate of heating and cooling. Generally, dry surfaces heat and cool faster than moist surfaces. Plowed fields, sandy beaches, and paved roads become hotter than surrounding meadows and wooded areas. During the day, air is warmer over a plowed field than over a forest or swamp; during the night the situation is reversed.

The distribution of water vapor and cloudiness is another important factor influencing air temperature. Although areas with a high percentage of cloudiness have a high degree of reflectivity, the energy which is not reflected is easily trapped in the lower layers due to the greenhouse effect. Thus, it must be expected that areas of high annual moisture content will have relatively high annual temperature.

Air Circulation

The higher mean temperature of the Northern Hemisphere is not only an effect of its greater land cover, but the oceans are also warmer than in the Southern Hemisphere. This is partly due to the movement of warm equatorial waters from the Southern Hemisphere into the Northern Hemisphere, caused by the southeast trades which cross the Equator. Another factor conducive to higher mean temperatures in the Northern Hemisphere is the partial protection of its oceans from cold polar waters and Arctic ice by land barriers. There is no such barrier between the Antarctic region and the southern oceans.

TOPOGRAPHY

Over land, climates may vary radically within very short distances because of altitudes and the variations in land forms. In this section we will discuss these two general effects.

Altitude

The height of an area above sea level exerts a considerable influence of its climate. For instance, a place located on the Equator in the high Andes of South America would have a climate quite different from a place located a few feet above sea level at the same latitude.

A change of all climatic values is observed as a function of elevation.

Land Forms

A powerful influence on climates is mountainous terrain, especially the long high chains of mountains that act as climatic divides. These obstacles deflect the tracks of cyclones and block the passage of air masses in the lower levels. If the pressure gradients are strong enough to force the air masses over the mountains, the forced ascent and descent will modify the air masses to a great extent, thus modifying the climate on both the windward and leeward sides.

The alinement of the mountain range may block certain air masses and keep them from getting to the lee side of the mountains. For example, the Himalayas and the Alps, with an east-west orientation, prevent fresh polar air masses from advancing southward. Therefore, the climates of India and Italy are warmer in winter than other locations of the same latitude. The coastal ranges of mountains in North America, running in a north-south line, prevent the passage of unmodified maritime air masses to the lee side.

Probably the most noted effect of mountains is the distribution of precipitation. The precipitation values, level for level, are much higher on the windward side.

In regions where the prevailing circulation flows against a mountain barrier, the amounts of precipitation increase more or less uniformly toward the tops of the mountains. This occurs on the windward side of the mountains to elevations of about 10,000 feet. However, in the trade wind zone, such as at the Hawaiian Islands, precipitation amounts increase only to about 3,000 feet and then decrease gradually. Even with this decrease in amounts, more rain is received at 6,000 feet than at sea level.

OCEAN CURRENTS

The currents of the oceans have considerable influence on the climate of certain areas and must be considered when preparing operational forecasts. However, since a discussion of ocean currents is presented in chapter 22 of AG 3 & 2, NT 10363-D, they are not discussed in this manual.

OTHER FACTORS INFLUENCING CLIMATE

Vegetation and human activity, though their roles may be minor in the overall climatic picture, do have an effect, at least in a local area, on the climate of that locality.

It is believed that large human settlements and industrial plants have a large influence on climate. The atmospheric pollution is increased and the radiation balance in the vicinity of the pollution is changed. This affects the daily maximum and minimum temperatures inside cities. They are generally higher than in the suburbs. The higher concentration of hygroscopic condensation unclei in the cities results in an increased number of fogs. Too, with the larger heat source concentrated over cities, increased convection gives rise to greater amounts of cloudiness with slightly higher amounts and frequencies of rain.

Evaporation and transpiration (breathing) of plants are very important influences on climate. Falling precipitation caught in trees before reaching the ground may be evaporated. Precipitation which reaches the ground does not readily evaporate, nor run off easily, due to the spongy structure of solid forests that can absorb and store considerable quantities of water. Snow in forests is protected from direct radiation by the trees and may stay on the ground for much longer periods than over open, exposed surfaces. Inside forests temperature maximums and minimums are higher than over open land at the same latitude. Relative humidities are higher in forests and wind speeds are considerably lower than outside.

CLIMATOLOGICAL DATA

Climatological records are based on the meteorological observations that are taken at a particular locality. This information may be presented in a number of ways.

Temperature records generally include the following temperature values: Daily maximums and minimums by months; the extremes, the average temperature by year and month; the mean monthly and annual temperature; and the mean monthly maximum and minimum. These values or elements were discussed earlier in this chapter. Sometimes included are the monthly and seasonal degree days. A degree day is the departure of the mean temperature from a daily average temperature of 65°F. For instance, if the mean temperature for a particular date were 50°F., the number of degree days for that date is 15. Of great climatic significance is the range between the mean temperature of the warmest month and the coldest month. Other temperature data are sometimes given. These may include the number of days with the following temperatures: Maximum of 90°F and above; maximum of 32°F and below; minimum of 32°F and below; and minimum of 0°F and below.

Precipitation records include the mean annual and monthly totals. The range between the highest and the lowest annual rainfall for a locality is the best indication of the dependability of the precipitation. The records often show the absolute maximum rainfall and snowfall for a 24-hour period by months, as well as the maximum and minimum precipitation for each month.

Climatic records usually show data on winds. Such information indicates the mean hourly speed and the prevailing direction by month. Also shown are the speed and direction of the fastest mile of the wind for the 12 months and the year in which it occured.

Data on cloudiness, humidity, thunderstorms, and heavy fog are often included. Other helphful data would be the frequency and distribution of cyclones and anticyclones; passage of fronts; the proportion of rainfall and snowfall received from cyclonic storms and local, airmass thunderstorms; and climatological data on upper air conditions.

METHODS OF PRESENTATION

Climatological information is presented in many different ways. Tables are frequently used. Maps are particularly useful in presenting climatic information in cases where geography is an important factor. Wind data can often be given by means of a device called a wind rose, which presents information on the prevailing wind directions. For an example of a wind rose, see figure 2-4.

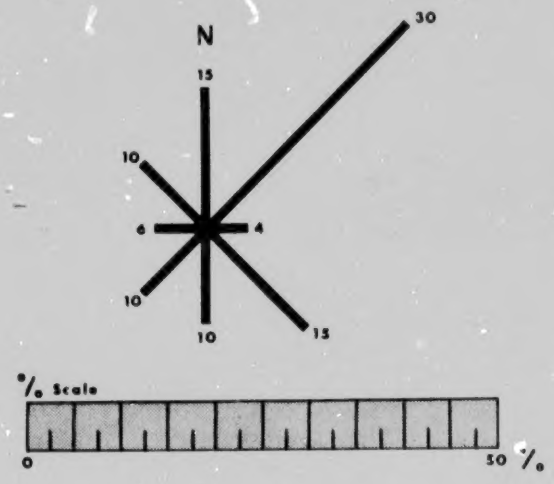

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Figure 2-4.—A wind rose.

Graphs are usually divided into bar and line graphs, or the graph may be a combination of the two. Figure 2-5 shows an example of a bar graph and a line graph of the same information.

AVAILABILITY OF DATA

Every naval weather service office should have climatological records in usable form available for the area in which the office is located, and for such other areas as may be necessary to carry out its mission. Various climatological records that are available for Naval Weather Service units from the NWSED, National Climatic Center, Asheville, NOrth Carolina include: the Summary of Meteorological Observations, Surface

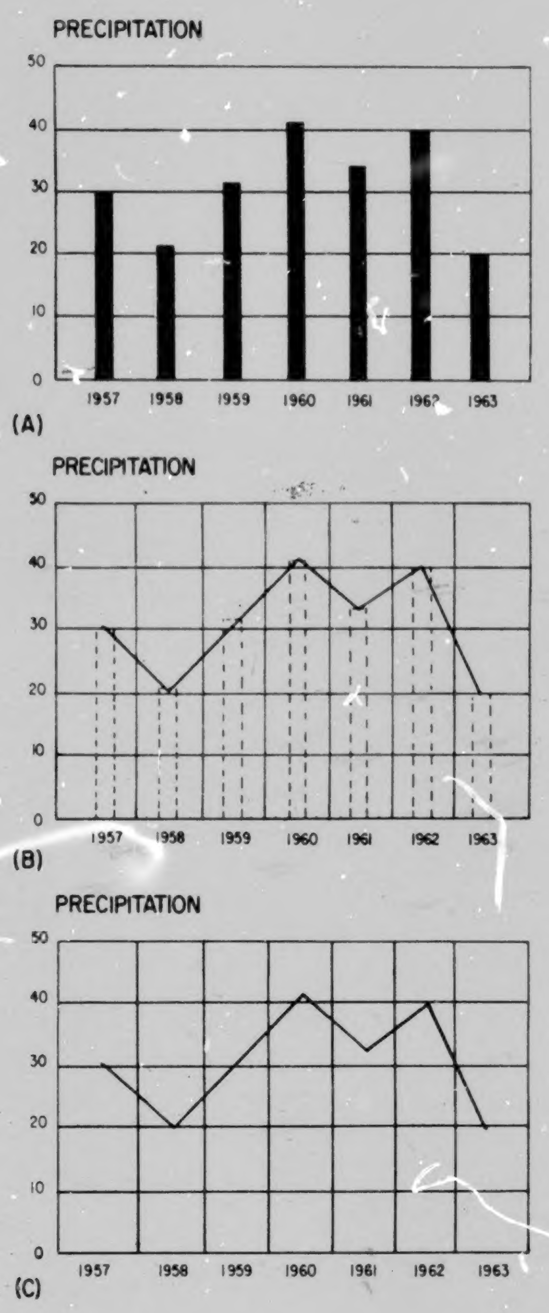

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Figure 2-5.—A comparison of the bar and line graph methods of showing the variable annual precipitation in a time series. (A) Bar graph; (B and C) line graphs.

(SMOS); Local Climatological Data (LCD) for selected stations; Cross-Wind Summary; Summary of Synoptic Meteorological Observations (SSMO); and Winds Aloft Summaries.

The first two are routinely issued and revised; the last three are prepared and issued only on request.

Frequency SMOS

These standard summaries are prepared from Navy Monthly Meteorological Records (MMR). Each SMOS is for a specific station. Frequency distributions for various parameters are presented by time of day, month, and year. Each SMOS is revised every 5 years.

Local Climatological Data

The Local Climatological Data summary consists of means and extremes (temperature, precipitation, wind, etc.) by month, mean temperature and total precipitation by months for specific years of record, and monthly and seasonal degree days. It is revised annually.

Cross-Wind Summary

This summary presents the precentage of occurrence of cross-winds for a given runway.

Summary of Synoptic Meteorological Observations

These standard summaries are similar to the Summary of Meteorological Observations, Surface (SMOS) in that they present a number of useful monthly and annual tabulations of surface climatological data and various combinations of the included parameters. They are specifically designed for compatibility with the various weather reporting codes of foreign nations and of ships at sea.

Winds Aloft Summary

These summaries are either seasonal or monthly and are generally prepared for various constant height levels and constant pressure levels. The summaries contain winds aloft data giving speed and directions over the period covered. The legend on each chart is self-explanatory.

Worldwide Airfield Summaries

These summaries provide climatological data for airfields and geographical areas throughout the world. There are 10 volumes, some published in two or more parts.

The summaries are prepared at the National Climatic Center (NCC) Asheville, North Carolina. The Headquarters, Naval Weather Service Command, controls the Navy participation in the joint activity of the Asheville operation. Requests for these summaries should be forwarded via official channels to:

Commander, Naval Weather Service Command
Washington Navy Yard
Building 200
Washington, D. C. 20390

with a copy to the Officer in Charge, NWSED, NCC, Asheville, North Carolina.

CLIMATOLOGICAL REFERENCES

There are many references which can be used in climatological work, so many in fact that they would be too numerous to list all of them here. Most of the Navy climatological references are listed in the Navy Stock List of Forms and Publications, NavSup Publication 2002, Cognizance Symbol I, Part C, Section VIII. Navy Climatological publications are found under the NA/NW 50-1C series. A few of those available are listed as follows:

1. Guide to Standard Weather Summaries, NavAir 50-1C-534. This publication contains an index of all the standard machine-tabulated summaries available and on file at the National Climatic Center.

2. U. S. Navy Marine Climatic Atlas of the World, Vols. I through VIII. These publications contain climatic data for all the principal ocean areas of the world. It has both a surface and an upper air section. The surface section contains data presented by graphs, tables, and isopleths on such elements as surface winds, gales, visibility, precipitation, etc. It also includes charts containing isobars of normal pressure and principal storm tracks. The upper air section includes charts of upper wind graphs, temperatures, relative humidity, and heights for selected levels; tropopause data; and refractive index gradients.

3. The local Area Forecaster's Handbook. This handbook as required by NAVWEASERCOM Instruction 3140.2( ) contains valuable information on local and area weather as follows: A description of the local topography and terrain, general synoptic characteristics of weather occurrences in the local area, mean storm tracks for your region, a limited amount of climatological data, and local forecasting rules or techniques. It can serve as a composite summary of expected weather events and the effect of certain parameters on the local weather.

Stations that are preparing their first handbook or revising their old one can make excellent use of their Summary of Meteorological Observations, Surface (SMOS). Individual tables or graphs thereof may be placed in the handbook, or the entire SMOS may be used as a separate appendix to supplement the descriptive text. Extra copies of SMOS are available from the Headquarters, Naval Weather Service Command for this purpose. If the SMOS is used as a separate appendix, it need not be forwarded with the handbook to the Headquarters, Naval Weather Service Command. Many excellent suggestions for climatological preparation can be found in NW 50-1C-536, Climatology at Work. This is also an excellent statistics primer.

The handbook, to be of the utmost use and value to the station forecasters, must be updated and revised periodically.

CLIMATOLOGICAL SERVICES

The Naval Weather Service Command provides climatological services to the Navy through its weather units, Fleet Weather Facilities and Centrals, and the NWSED, NCC, Asheville.

Weather units at shore stations must provide climatological studies for their local areas based upon locally available weather data and climatological publications. Requests for climatological studies beyond the capabilities of the local weather unit should be submitted to the nearest Fleet Weather Central.

Fleet Weather Centrals provide, in addition to local area studies, climatological studies for their areas of responsibility, climatological studies for other weather units and forces afloat, including the preparation of the climatological portion of weather annexes for operation plans when requested. They also conduct active research in applied climatology.

INTERPRETATION

Climatological records are almost worthless unless they are interpreted correctly. Proper interpretation requires that all of the meteorological elements be studied so as to present a composite picture. One meteorological element alone means very little to a meteorologist. For instance, it is possible to conclude that Cairo, Egypt, and Galveston, Texas, have about the same kind of weather from a study of only the temperature, since the yearly and monthly means and annual range are approximately the same. However, Galveston has about 40 times as much precipitation. Thus, their weather conditions over the year must differ greatly.

To properly interpret just one meteorological element requires a study of several factors. The temperature of a particular locality must be studied from the stanpoint not only of the mean but also of the extremes, and the diurnal and annual ranges. The effectiveness of precipitation depends on several factors, such as amount, distribution, and evaporation. The mean precipitation for a particular month for a locality may be several inches, but the interpreter may find from a study of the locality's records that in some years the precipitation for that month is less than an inch, possibly not even a trace.

APPLICATION OF CLIMATOLOGY TO WEATHER PREDICTION

Climatology is introduced where operational planning is required for a length of time beyond the range covered by weather forecasting techniques. A study of the climate of that area or region may well foretell the general weather pattern to be expected.

A more direct application of climatology can be made by both the experienced and the inexperienced forecaster. The experienced forecaster who has personal experience at a particular station can use climatology as a refresher for the overall weather patterns which can be expected for the ensuing season. This knowledge can help him to be more perceptive in his everyday analyses, to be alert for changing patterns with the seasons, and to produce a higher quality forecast.

The inexperienced forecaster who has had no experience at a particular station must rely on climatology as a substitute for this experience.

Forecasters cannot be expected to become familiar overnight with the weather peculiarities of their new area of responsibility. The station indoctrination period can be greatly reduced if the new forecaster is furnished with "packaged experience" in a form which can place him more nearly on a par with those forecasters already experienced at that station. The Local Area Forecaster's Handbook is a good example.

USES AND LIMITATIONS OF CLIMATOLOGICAL DATA

The Naval Weather Service makes many uses of climatological data. In using the data, however, it must be clear that climatology has its limitations in the field of meteorology. It may be put this way. Climatology is an essential supplement to meteorology, but it must never be considered a substitute for the meteorological situation which constitutes current weather conditions.

The Navy makes use of climatological research in determining the locations for naval bases and naval air stations. The prevalence of fogs, the prevailing wind directions, the occurrence of low ceilings, and the passage of fronts, along with the frequency of thunderstorms, are some of the important considerations in establishing naval air stations in particular.

In the locating of naval air stations, fogs are especially important as a consideration. Ground fog will tend to drain into an area that is located in a topographical depression. When ground fog occurs, a station located at such a site would be the first to be fogged in, and it would be the last to have fog dissipation. If advection fog is predominant in the area, its effect can be minimized by selecting a site where the prevailing wind during such fogs has a downhill component so that adiabatic compression, with its associated temperature rise, will dissipate fog.

If the air station is to be located in an industrial area, it should be located to the windward side of the industrial area to minimize the reduction of visibility due to smoke. Air stations should not be located near the lee side of high obstructions which could produce troublesome eddy currents.

Also, in the establishment of a naval air station the occurrence of low ceiling must be considered. A study of the climatological data of an area will reveal the prevalence of ceilings which would adversely affect flight operations.

The passage of fronts is another important consideration. It is very essential to know the frequency of frontal passage, the normal speed of the movements of fronts, seasonal fronts, and the usual weather conditions that accompany the different types of fronts that move across the area.

The distribution and frequency of thunderstorms are important climatically in the location of naval air stations. This is due to the great destructibility of gusty winds accompanying thunderstorms, and the disruptions that they can cause in the flight program.

The orientation of the runways at an air station can be properly made only after due consideration of the prevailing winds of the area. Prevailing winds must be studied carefully. It is necessary to consider several directions that occur frequently. Wind directions associated with low wind speeds may be somewhat ignored if the speed is low enough that crosswind landings may be made safely.

In using climatological records for study, it is necessary to consider all elements and values that are available. It is a mistake to rely too heavily on the normal values that are attached to certain weather elements, since one year or month may vary a great deal from the normal for the year or month.

WORLD WEATHER

Factors to be considered in analyzing weather in different areas of the world include world air masses and fronts, the polar front, regional circulations, oceanic weather, and arctic and antarctic weather. These are discussed in the following paragraphs, concluding with a discussion of United States weather as it relates to world weather.

WORLD AIR MASSES AND FRONTS

The initial characteristics of an air mass is determinded by the surface (land or water) over which the air mass originates (its source region). However, these initial characteristics undergo a variety of changes once the air mass begins to move out of the source region.

The regions of high pressure from which principal airflows originate are the primary source regions for the air masses that move over the world. The relative intensity of the continental and maritime highs determines the location of the zones of convergence between them. In these zones of convergence, where air masses with different properties meet, lie the polar fronts. The transient low-pressure cells of the extratropical cyclone move along the polar fronts.

Air-Mass Source Regions and Classifications

Figures 2-6 and 2-7 show the world air masses, fronts, and centers of major pressure systems in January and July, respectively.

A more detailed discussion of air mass source regions and classifications is presented in AG 3 & 2, Chapter 6, NT 10363-D.

Air-Mass Characteristics

The weather characteristics of a particular month in a given locality are governed by:

1. The intensity of the airstream flowing over that particular region.
2. The origin of the air mass within the stream.
3. The previous history of the air mass before arriving over the region.
4. The effects of local topography upon the several meteorological processes.
5. The proximity of the region to one of the zones of atmospheric convergence.

The rate of transformation depends upon the degree of difference in temperature and moisture, the speed of airflow, and the amount of

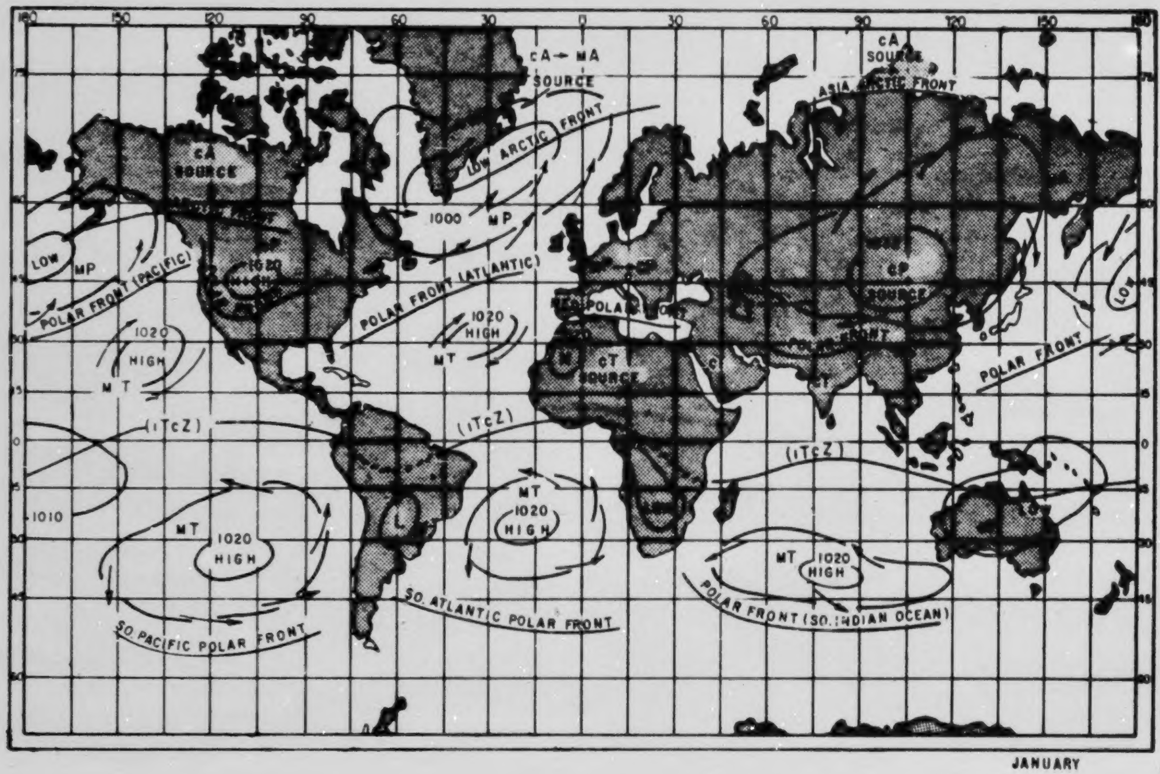
cA → MA
SOURCE
LOW ARCTIC FRONT
cA
SOURCE
1000
MP
LOW
MP
POLAR FRONT (PACIFIC)
1020
HIGH
POLAR FRONT (ATLANTIC)
1020
HIGH
MT
1020
HIGH
MT
(ITcZ)
(ITcZ)
1010
MT
1020
HIGH
MT
1020
HIGH
SO. PACIFIC POLAR FRONT
SO. ATLANTIC POLAR FRONT
cA
SOURCE
ASIA ARCTIC FRONT
cP
SOURCE
POLAR FRONT
cT
SOURCE
POLAR FRONT
LOW
(ITcZ)
MT
1020
HIGH
POLAR FRONT (SO. INDIAN OCEAN)
180
150
120
90
60
30
0
30
60
90
120
150
180
JANUARY

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Figure 2-6.—Chart showing world air masses, fronts and centers of major pressure systems in January.

turbulence in the lower layers. Discussion of the modification process is facilitated if the following three classes of conditions are considered.

1. Warm air moving over a colder surface. In this case, the surface layer of air undergoes rapid transformation, but, because of the stability produced by this surface cooling, the effects do not extend to great heights.

2. Cold air moving over a warmer surface. In such a situation, the surface layer of air undergoes less rapid modification, but the effects extend to much higher levels because of the turbulent and convective exchange of heat and moisture.

3. Warm, dry air moving over a warm, moist surface. In this case, the air mass picks up moisture rapidly, but, because of the small amount of heating at the surface, the effects reach only a limited height. Under such conditions, a shallow layer of unstable air exists at the surface with a stable inversion layer above. This prevents the distribution of moisture to high levels.

POLAR FRONT

There are four principal types of air masses in each hemisphere, named according to their source regions: arctic, polar, tropical, and equatorial. Arctic air is separated from polar air by the Arctic front. The polar front exists between tropical and polar air masses. The ITCZ lies in the equatorial air mass region.

The polar front does not constitute a single, more or less continuous, band of convergence and bad weather. It appears, rather, as a series of

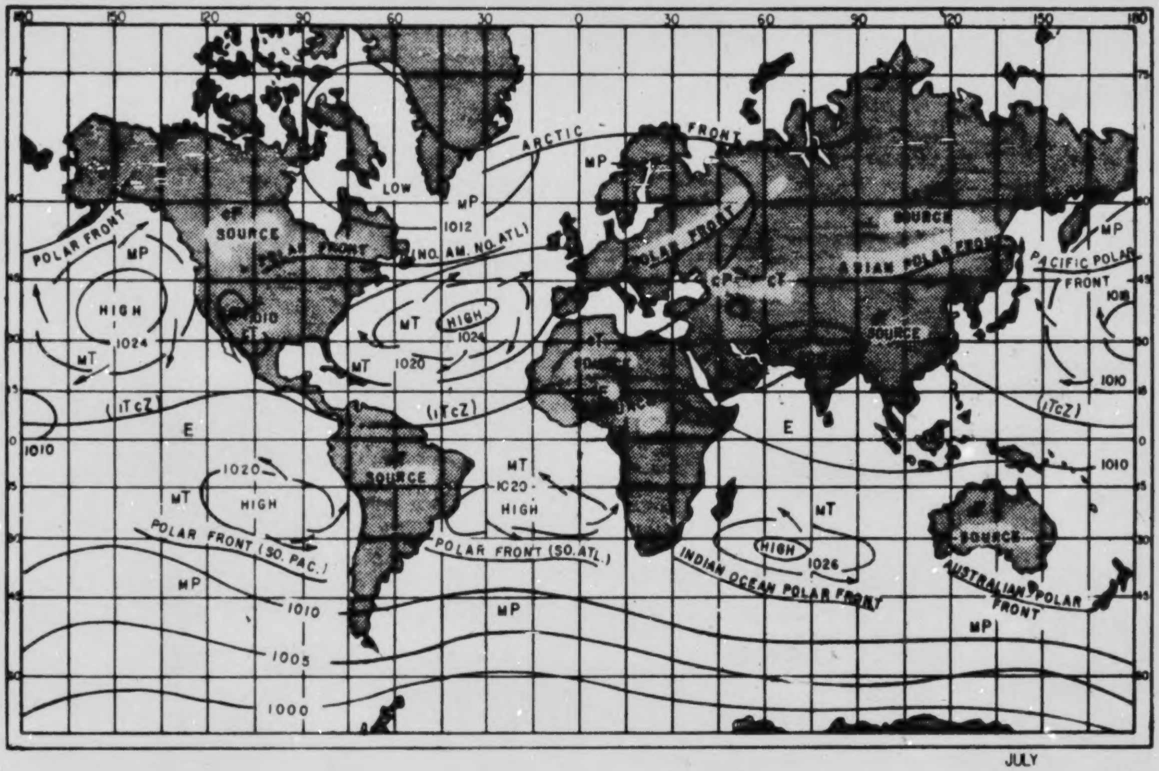
180
150
120
90
60
30
0
30
60
90
120
150
180
LOW
MP
1012
ARCTIC FRONT
MP
cP SOURCE
POLAR FRONT
POLAR FRONT
(NO. AM. NO. ATL.)
MP
HIGH
1024
MT
1010
cT
MT
HIGH
1024
MT
1020
POLAR FRONT
cP
cT
SOURCE
ASIAN POLAR FRONT
SOURCE
MP
PACIFIC POLAR FRONT
1010
1010
(ITCZ)
(ITCZ)
(ITCZ)
E
E
1010
1010
1020
HIGH
MT
SOURCE
MT
1020
HIGH
POLAR FRONT (SO. PAC.)
POLAR FRONT (SO. ATL.)
MT
HIGH
1026
INDIAN OCEAN POLAR FRONT
SOURCE
AUSTRALIAN POLAR FRONT
MP
MP
MP
1010
1005
1000
JULY

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Figure 2-7.—Chart showing world air masses, fronts, and centers of major pressure systems in July.

active, convergent zones, which move in a general direction from west to east. These convergent zones are the migratory low or traveling cyclones of the middle latitudes. It has been shown that, in middle and high latitudes, it is impossible for air masses with different densities (temperatures) to lie side by side continuously in equilibrium, but that waves will form along the discontinuity or front that separates the two. If the wavelength of the disturbances are of the order of 300 to 1,800 miles, the waves are said to be unstable and, their amplitude will increase with time. In this case, the cold air in the rear of the waves is pushed toward the Equator, where it tends to underrun the warm air, thus forming the cold front of the cyclone. A compensating polarward flow of warm air occurs in the front portion of the advancing wave, thus forming the warm front of the cyclone. The warm, light air behind the warm front tends to ride up and over the denser cold air ahead of the front. The waves develop into occlusions, and a typical extratropical cyclone is the result.

The formation of a cyclone along the polar front is first noted as a slight indentation of the front. This indentation or wave deepens and moves along the front, rapidly at first, and then becomes progressively slower. The waves nearly always move from west to east for two reasons. In the first place, this is the prevailing direction of the windstream in which they form. Secondly, according to wave theory, a wave type cyclone should move in such a way that the warmer air is on the right in the Northern Hemisphere and on the left in the Southern

Hemisphere. Since the warmer air is usually to the south of the polar front in the Northern Hemisphere, and to the north of the polar front in the Southern Hemisphere, the wave movement also contributes to the west-to-east motion of cyclones. Thus, the zones of active convergence along the polar fronts in each hemisphere appear as migratory lows which travel from west to east, generally in the region which is more than 20° from the Equator.

REGIONAL CIRCULATIONS

Many regions have unusual local weather phenomena caused directly by the temperature difference between land and water or by local topographical features. Knowledge of the regional circulations which have significant effects on operating conditions is essential for naval operations.

Land and Sea Breezes

Ocean breezes generally occur on a summer afternoon near a coastline. When an air mass is stationary over a coastline and the land surface is warmer than the sea surface as a result of radiation heating, an onshore breeze is produced at the surface, with an outward drift aloft. At night, the radiation cooling of the land is greater than that of the sea; the horizontal temperature gradient is, therefore, normally reversed. This produces the offshore breeze, which reaches its maximum about dawn. The land and sea breeze phenomenon is purely local. Its influence does not extend far from the coastline and seldom exceeds a height of 3,000 feet.

Monsoon Winds

Monsoon winds are characterized by a tendency toward a reversal in prevailing wind direction between winter and summer.

Because of the great size of the continent, the monsoon is most developed over eastern and southern Asia. However, monsoons in modified form, or monsoon tendencies are characteristic of other regions as well. Southeastern United States, northern Australia, Spain, and South Africa are regions with monsoon tendencies. These land areas are not extensive enough to cause a complete seasonal reversal of winds, as does Asia, but they produce partial monsoons, which have a significant effect on the amount of seasonal rainfall.

Regions with strong monsoon tendencies usually are on the eastern sides of continents. This is especially true in the middle latitudes, since the western or windward coasts are distinctly marine in character, with only small changes in temperature from winter to summer. It is, therefore, only on the more continental eastern, or leeward, sides that sufficiently large seasonal extremes of temperature can develop to produce a wind reversal.

Monsoon air, as observed over India and Burma, consists of unmodified equatorial air during the summer monsoon and of highly modified cP air during the winter season. The weather conditions during the summer monsoon consist of cloudy weather with almost continuous rain and widespread shower activity. High temperatures and humidities are the rule. Weather conditions during the winter monsoon are dominated by the dry, adiabatically warmed polar air flowing toward the Equator. It is during this time that generally pleasant weather prevails over most of the area influenced by monsoon conditions.

The reverse of this condition occurs along the east coast of Indochina, particularly the northern coastal sections of Vietnam. During the winter the cP air flowing southward over the East and South China Seas becomes sufficiently modified to produce extensive cloudiness and drizzle to this area. This phenomenon is called the "Crachin."

Ravine Winds

CHARACTERISTICS.–Strong winds often blow along deep valleys or ravines that break across mountain ranges. Some of the highest observed wind velocities have occurred in such valleys. The mistral of southern France and the bora of Trieste (located on the Adriatic Sea in northeastern Italy) are among the best known of a large group of such winds. All ravine winds are characterized by their violence, and most of them have a marked seasonal variation. It is believed that ravine winds are due to a pressure gradient acting along the ravine. The wind blows directly from high pressure to low, along the ravine; the wind speed increases with distance from the source, reaching a maximum as it emerges at the low-pressure end of the ravine.

The wind occurs more frequently in winter than in summer. It is strongest and most frequent in the early morning.

RAVINE WINDS OF THE UNITED STATES.–Many examples of ravine winds are found in the Pacific Coast Ranges and the coastal areas of the United States. The best known include the following:

1. The gales in the Columbia River Gorge across the Cascades.
2. The southeasterly winds through the San Joaquin Valley in California.
3. The Santa Ana wind in the Los Angeles region. The Santa Ana blows through the Cajon Pass between the San Gabriel and the San Bernardino ranges. At the high-pressure end over the Mojave Desert, speeds are low, but 15 miles farther south, speeds often exceeds 50 mph. At the low-pressure end, it may exceed 60 mph for many hours. The Santa Ana blows only in the winter half of the year, as the temperature discontinuity responsible for the pressure gradient along the Cajon Pass is absent at other times. This is generally true of the Pacific Coast ravine winds.

OCEANIC WEATHER

Since the naval vessels of the United States now operate regularly in virtually all the oceanic areas of the earth, you as Aerographer's Mates must be acquainted with oceanic weather. Some general considerations of the weather encountered over ocean areas are given in this section.

Because land and water heat and cool at different rates, the location of continents and oceans greatly affects the earth's pattern of air temperature and therefore influences the weather.

Since the upper layers of the ocean are nearly always in a state of motion, heat loss or heat gains occurring at the sea surface are distributed throughout large volumes of water. This mixing process sharply reduces the temperature contrasts between day and night and between winter and summer.

Oceanic Weather Control

It has been long recognized that the influence of the ocean plays an important part in climate and weather, particularly in the realm of temperature, humidity, and precipitation. This is only natural, since three-fourths of the earth's surface is covered by water.

The two climatic extremes that relate to water and land distribution over the earth are MARITIME and CONTINENTAL. Continental climate is generally evidenced by a wide range in annual and diurnal temperatures, little cloudiness, and little precipitation. Continental climate is a product of a minimum influence by the oceans. Maritime climate, on the other hand, prevails over the oceans, which spawn it. Characteristically its temperature range both annual and diurnal, is small and its precipitation and cloudiness are great. The great inland areas of North America and Eurasia, for instance, show the effect of being a great distance from the oceans. Much of the moisture has been precipitated by the time the air masses have moved to the interiors from the oceans, leaving little for those areas. The humidity is less over areas away from oceanic influence, allowing more of the sun's insolation to reach the earth's surface and causing the earth's surface to lose more of its heat through terrestrial radiation; these processes help to cause a large diurnal range in temperature. In the summer the relatively cool air masses from the oceans on the west coast of the continents are warmed as they move inland and lose their maritime effect. On the other hand, the relatively warm air masses in the winter are cooled. The northwest coastal area of the United States shows the effect of the maritime influences of the Pacific Ocean. The ocean brings to that region a small daily and annual range in temperature, high humidity, and abundant rainfall.

Water vapor is considered one of the most important variables in meteorology. The state of the weather is largely expressed in terms of the amount of water vapor present and what is happening to the water vapor. Two principal elements of climate, which are precipitation and humidity, are dependent upon water vapor. The oceans are the main source of water vapor. Thus, Aerographer's Mates should see readily that the weather is largely controlled by the oceans.

Effects of Air-Sea Interchange

The atmosphere and the oceans have tremendous effects on each other. These effects are principally in the realm of temperature and water vapor. For a fuller picture of the interchange of the atmosphere and the oceans, you as Aerographer's Mates must consider the various processes that control the heat balance of the air and sea.

The heat balance of the oceans is maintained by the processes of radiation, the exchange of sensible heat, and the evaporation and condensation of water vapor on the sea surface.

The amount of radiant energy absorbed by the sea depends upon the amount of energy reaching the surface and the amount of reflection. When the sun is directly overhead, the amount of its energy that is reflected amounts to only about 3 percent. Even when the sun is 30° above the horizon, the amount of reflection is just 6 percent. There is a reflection of about 25 percent of the energy when the sun is 10° above the horizon. (See fig. 2-8.) The amount of radiation which penetrates the sea surface depends upon the reflection loss from the total amount of incoming radiation. Reflection loss is especially great in the presence of waves when the sun is low.

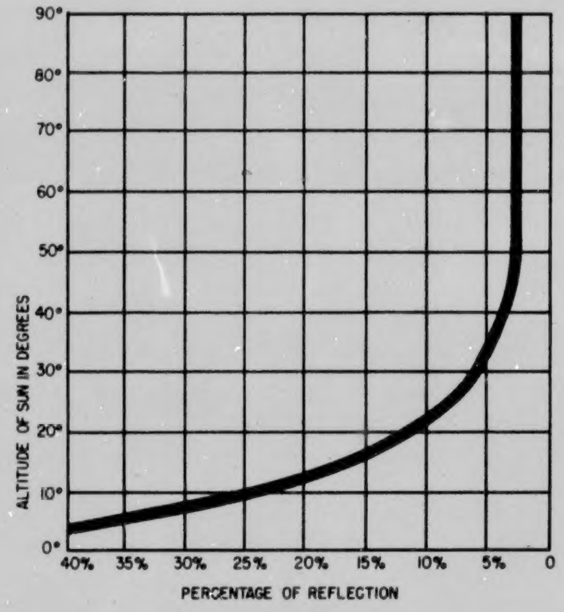

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Figure 2-8.—Percentage of reflected radiation.

Most of the energy is absorbed in the first meter of sea water. This is true of the clearest water as well as of quite turbid water. In water that is extremely turbid the absorption is in the very uppermost layers. Foam and air bubbles are two major causes of a proportionately greater amount of absorption in the uppermost meter of the sea. However, due to vertical mixing the heat absorbed in the upper layer is carried to great depths of the ocean, which acts as a great storage reservoir of heat.

There is an exchange of energy between the oceans and the atmosphere. The surface of the oceans emits long-wave heat radiation, the energy of this radiation being proportional to the fourth power of the surface's absolute temperature. The sea surface at the same time receives long wave radiation from the atmosphere. Although some of this incoming radiation from the atmosphere is reflected from the surface of the oceans, most of it is absorbed in a very thin layer of the water surface, since at long wavelengths the absorption coefficients are great. The difference between the incoming long-wave atmospheric radiation and the outgoing long-wave radiation from the sea surface is known as the EFFECTIVE BACK RADIATION. The effective back radiation depends primarily on the temperature of the sea surface and the water vapor content of the atmosphere. The time of day and the season have little effect on effective back radiation, since the diurnal and annual variation of the sea surface temperature and of the relative humidity of the air above the oceans is slight.

In order for conduction to take place between the oceans and the atmophere, there must be a difference in the temperature between the ocean surface and the immediately overlying area. On the average the temperature of the surface of the oceans is higher than that of the overlying air. It might be expected that all of the ocean's surplus of heat is either radiated or conducted to the atmosphere. This is not the case. Only a small percentage of the ocean's surplus heat is actually conducted to the atmosphere. About 90 percent of the surplus is used for evaporation of ocean water.

Due to the processes of radiation and mixing, the oceans act as a thermostat relative to the atmosphere. The energy stored at one place during one season may be given-off at another locality and during a later season. Hence, there seems to be a constant effort by the atmosphere and the oceans to keep their temperature in balance by an interchange of heat.

STABILITY.–As you know, the deciding factor of most weather phenomena is the stability of the atmosphere. Air masses may become more stable or less stable as they move over ocean surfaces. The temperature contrast between the ocean surface and the lowest layers of the overlying air determines whether the ocean will cause stability or instability.

When the air moving over the ocean has a higher temperature than that of the ocean surface, the lower layers of the air become stable. As the air cools and hovers over the ocean, conduction of heat from the air to the ocean diminishes rapidly with the sharp reduction in the convective activity of the air.

On the other hand, when the air mass is colder than the ocean surface over which it is moving, there is resulting instability, as the colder air is warmed by the ocean, and convective activity increases. If the warming by the ocean is sufficiently intense, the convective currents may rise rapidly to such heights that thunderstorms develop.

MOISTURE CONTENT.–The interchange of moisture between the atmosphere and the oceans is one of the most important features of the whole meteorological picture. Without this interchange, weather as we know it could not exist; there would be no clouds and no precipitation. The oceans are by far the greatest source of moisture for the atmosphere. Other moisture sources are negligible in comparison.

Whether the atmosphere gives up some of its moisture to the ocean or vice versa, depends greatly upon the vapor pressures of the two. You may recall that vapor pressure is the pressure exerted by the molecules of water vapor in the atmosphere or over the surface of liquid water. When the vapor pressure of a liquid is equal to that of the atmosphere above the liquid, there is little or no apparent interchange of moisture. In other words, at equal vapor pressure, as many molecules escape from the liquid to the atmosphere as reenter the liquid from the atmosphere. This would be true of air that had become saturated. The saturation vapor pressure increases with increasing temperature.

If the temperature of the surface water of the ocean is warmer than that of the air, the vapor pressure of the water at its surface is greater than that of the air. When this condition exists, there can be abundant evaporation from the ocean surface. This evaporation will be aided by the turbulence of the air brought on by the unstable condition of the lowest layers. It follows, then, that the greatest evaporation takes place when cold air flows over warm ocean waters.

Let us consider the opposite condition–warm air flowing over a relatively cold body of water. When this happens, there is stable stratification in the lowest layers of the atmosphere. The vapor pressure of the air soon reaches a state of equilibrium with that of the water surface. Evaporation stops. However, if the warm air is quite moist, it is possible for the moisture in the air to condense on the water surface. Contact of the warm air with the cold water may result in the formation of fog by lowering the air temperature to the dewpoint.

The direct interchange of moisture from the atmosphere to the oceans is in the nature of precipitation and, to a minor extent, of condensation of moisture upon the ocean surface. The direct interchange, however, is not as meteorologically important as the indirect interchange. The indirect interchange is in the nature of the runoff from the land surfaces, which follows a sequence of events beginning with the evaporation of water from the ocean surfaces and ending with the subsequent condensation and precipitation over land areas. It may be thought that there is more precipitation over the oceans than over the land. This is not always the case, as the ocean surfaces do not favor precipitation nearly as much as land surfaces.

The relatively low evaporation in the tropical latitudes is attributed to the high relative humidities of the atmosphere and the low wind speeds of the equatorial belt. In the middle and high latitudes the lower relative humidities and higher wind speeds favor the evaporation from the warm surfaces to the relatively cold air.

The moisture supplied to the atmosphere in the higher latitudes varies greatly from winter to summer. During the summer months, the air over the sea surfaces is usually as warm as or warmer than the sea surfaces. In this case, evaporation is at a minimum. On the other hand, the air is colder than the ocean surfaces during the winter, and evaporation is at a maximum.

Equatorial and Tropical Weather

In the temperate zone, where westerly winds predominate, pressure patterns move in an easterly direction. In the tropics, however, weather moves in the opposite direction. The normal moist layer is 5,000 to 8,000 feet deep, rising above 12,000 feet during periods of unfavorable weather. Convergence occurs in opposing trade wind streams, northward flowing air, and cyclonic curvature. The deep, moist layer and the convergence account for the weather in equatorial and tropical regions.

Tropical cyclones, waves in the easterlies, the intertropical convergence zone, and other bad weather zones are discussed in chapter 12 of this training manual.

North Atlantic and North Pacific Oceans

POLAR FRONTAL ACTIVITY.–In winter, the most favorable conditions for vigorous frontal activity are concentrated along the east coasts of North America and Asia. Cold air masses from continental sources meet warm moist air from over the oceans. The warm ocean currents along these coasts greatly accentuate the frontal activity. The great difference in the temperature of the air masses, caused by the contrasting characteristics and proximity of their sources and the great amount of moisture that feeds into the air from the warm ocean currents, accounts for the intensity and persistence of these frontal zones off east coasts in winter. Modification of the air masses as they sweep eastward across the ocean causes the modified frontal activity of west coasts. See figures 2-6 and 2-7 for the location of the following frontal zones.

1. Polar Fronts in the Atlantic. In the Atlantic, the polar front of winter swings back and forth between the West Indies and the Great Lakes area, with a maximum of intensity when the front coincides with the coastline. Waves with cold and warm fronts form along this polar front and move northeastward along the front. Like all cyclonic waves, they develop low-pressure centers along the frontal trough. They may grow into severe disturbances and go through the usual stages of development: formation, growth, occlusion, and dissipation.

These cyclonic waves occur in families. Each family of waves is associated with a southward surge, or outbreak, of cold polar air. The polar front commonly extends approximately through the Great Lakes area. As the polar air advances, it pushes the front southward. The outbreak occurs, and polar air, joining the trade winds, spills equatorward.

There is no regular time interval for the large outbreaks of polar air, but the average period between them is about 5 1/2 days. Under average conditions, there are from three to six cyclonic waves on the polar front between each outbreak of polar air. The first of these usually travels along the front that lies farthest to the north. As the polar air accumulates north of the front, the front is pushed southward, and the last wave therefore follows a path that starts farther south than the path followed by the first wave. These families of polar front cyclones appear most regularly over the North Atlantic and North Pacific in winter.

During the summer months, the polar front of the Atlantic recedes to a location near the Great Lakes region, with the average summer storm track extending from the St. Lawrence Valley, across Newfoundland, and on toward Iceland. Polar outbreaks, with their accompanying family groupings of cyclones, are very irregular in summer and often do not exist at all. Frontal activity is more vigorous in winter than in summer because the polar and tropical air masses have greater temperature contrasts in winter, and polar highs reach maximum development in winter. Both of these factors increase the speed of winds pouring into fronts. Over oceans of middle latitudes, a third factor helps to make winter fronts more vigorous than summer fronts. In winter, continental air becomes very unstable when it moves over the comparatively warm ocean surface; in summer, it remains relatively stable over the comparatively cool ocean. Summer frontal activity (in middle latitudes) is therefore weak over oceans as well as over land. The high moisture content of maritime air causes much cloudiness, but this moisture adds little energy to frontal activity in the relatively stable summer air.

2. The polar front activity of the Pacific is similar to that of the Atlantic, except that in winter there are usually two fronts. When one high dominates the subtropical Pacific in the winter season, the Pacific polar front forms near the Asiatic coast. This front gets its energy from the temperature contrast between cold northerly monsoon winds and the tropical maritime air masses they meet, and from the warm, moist Japan Current. In moving along this polar front of the Asiatic North Pacific in winter, storms occlude before reaching the Aleutian Islands or the Gulf of Alaska. Because of its steady cyclonic circulation, the Aleutian low becomes a focal center, or a gathering point, for cyclones. The occluded fronts move around its southern side like wheel spokes. This frontal movement is limited to the southern side of the Aleutian low because mountains and the North American winter high-pressure center prevent fronts from passing northward through Alaska without considerable modification.

When cold season cyclones reach the Aleutians and the Gulf of Alaska, cold continental winds from the Arctic feed them from the north, and cool mP winds from the ocean feed them from the south. Here, where Arctic air meets maritime air over relatively warm water, is the Pacific Arctic front of winter. Although many occluded storms dissipate in the Gulf of Alaska, others become strongly regenerated with waves developing on the occluded fronts.

When the Pacific subtropical high divides into two cells or segments (as it does 50 percent of the time in winter and 25 percent of the time in summer), a front forms in the vicinity of the State of Hawaii. Along this front, the Kona storms develop and move northeastward. Those that succeed in moving beyond the realm of the northeast trades, which stunt them, may develop quite vigorously and advance to the American coast, generally occluding against the mountains. When this second polar front exists, two systems of cyclonic disturbances move across the Pacific. Because of their greater sources of energy, however, storms that originate over the Japan Current and move toward the Aleutians are always more severe. In the Atlantic, a second polar front, similar in nature and cause to the second polar front of the Pacific, sometimes—though rarely—develops.

During the summer months, the Pacific polar front lies to the north of Kamchatka and the Aleutians and shows no rhythmic polar outbreaks.

AIR-MASS WEATHER.–Flying weather is usually at its best in tropical maritime air at its source, within the subtropical highs. Scattered cumulus and then patches of stratocumulus clouds may develop, but the sky is almost never overcast and the scant precipitation falls in scattered showers. Variable, mild winds prevail.

The excellent flying weather of these mT source regions commonly extends through the moving air masses some distance from the sources. Cloudiness in the mT air increases with increase in distance from the source. On flights from Hawaii, or from the Azores northward through northward-moving mT air, stratiform clouds increase to the north. On flights from Hawaii or the Azores southward through southward-moving mT air (or the northeast trades), cumuliform clouds increase. (We are here considering only Northern Hemisphere situations. A comparable pattern exists in the Southern Hemisphere.) (See figure 2-9 for winds and stability conditions around the subtropical high.)

A typical breakdown of the weather conditions which may be encountered in air masses around the subtropical highs is as follows:

1. North of a Subtropical High. Any mT air that moves northward becomes chilled over the cool ocean surface. A stratus overcast may form, and drizzle may fall. Farther north, low ceilings (usually below 1,000 feet) may reach the surface, producing fog. The mT air surges farthest north in summer, for then subtropical highs are best developed and polar fronts lie farthest north. This mT air brings most of the summer fogginess to northern seas and coasts. It brings

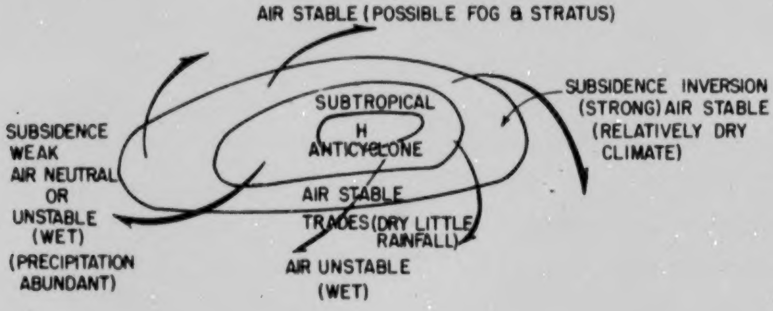
AIR STABLE (POSSIBLE FOG & STRATUS)
SUBSIDENCE INVERSION (STRONG) AIR STABLE (RELATIVELY DRY CLIMATE)
SUBTROPICAL
H
ANTICYCLONE
SUBSIDENCE WEAK AIR NEUTRAL OR UNSTABLE (WET) (PRECIPITATION ABUNDANT)
AIR STABLE
TRADES (DRY LITTLE RAINFALL)
AIR UNSTABLE (WET)

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Figure 2-9.—Winds and stability conditions around the subtropical high.

greatest fogginess in the Atlantic where it blows from over the warm Gulf Stream to over the cold Labrador Current (near Newfoundland), and in the Pacific where it blows from over the warm Kuroshio Current to over the cold Oyashio Current (near Kamchatka).

2. East of a Subtropical High. Along the California coast, and along the Atlantic coast of North Africa, the mT air blows from the west and the northwest. This air tends to remain stable for the following reasons:

a. It is coming from the northern, cooler portion of the source region.

b. Its surface layers remain cool, because it moves over cold ocean currents.

c. Its upper portions warm adiabatically because of subsidence.

Throughout the year, airways are smooth. The skies are clear to partly cloudy. Clouds are generally patches of stratocumulus, and rain is rare. The chief flight hazard in this air is coastal fog, which often hides the California or European coastal land. Stratus and stratocumulus clouds may cause the sky to be overcast, develop low ceilings, and produce drizzle that reduces visibility.

3. South of a Subtropical High. Where the mT air moves southward or southwestward (as trade winds), its lower layers are warmed by the tropical ocean surface. This produces scattered cumulus. Near the Equator, after absorbing much moisture and being heated, this air may develop cumulonimbus.

4. West of a Subtropical High. This mT air blows from the east and the southeast. Since it flows over warm water all of the way, the air neither chills nor heats. Over the ocean near the Philippines (and near Florida and the West Indies), this trade wind brings good flying weather—clear, or with scattered cumulus clouds. When it moves over land, this warm, moist air becomes unstable and turbulent and is a source of thunderstorms. When it moves over cold land (for example, southeastern United States in winter), it becomes stable and produces stratus clouds or fog. Over cold ocean surfaces, such as the Sea of Japan and the Kamchatka and Labrador Currents, it develops the persistent low stratus and fogs characteristic of these areas.

ARCTIC AND ANTARCTIC WEATHER

Geographically we think of the Arctic zone as being north of 66.5° N lat and the antarctic zone south of the Antarctic Circle at 66.5° S lat. Due to increased operation and importance of these zones, it is incumbent upon the Aerographer's Mate to familiarize himself with the prevailing weather and peculiarities of these regions. Only a brief resume of the weather conditions of these zones is given here. Many recently published texts and research papers provide additional information on these zones.

Arctic Weather

Today the Arctic is rapidly becoming the aerial crossroads of the world. This is not only due to the shorter Arctic routes between some of the major cities of the world, but also because flying weather over the Arctic is generally better than that encountered over the familiar ocean routes. As we learn more about Arctic weather and its effects upon flying, Arctic flying will become even more common.

To understand some of the important problems of the Arctic, the broad, underlying causes of the Arctic climate must be understood.

SEASONAL TEMPERATURE VARIATIONS.—From our previous discussion of climatic controls we have seen that the most important factor that determines the climate of any area is the amount of energy it receives from the sun.

Look back to figure 2-2 and you will note that during the winter much of the Arctic receives little or no direct heat from the sun. In comparison notice the amount of sunlight received at different latitudes. The cold, winter temperatures common in the Arctic result from a lack of the sun's energy.

Since the energy from the sun is not the only factor responsible for the climate, we must consider two other factors: the land-sea-ice distribution and mountain barriers. All three contribute to the tremendous variation in climate at different points at the same latitude in the Arctic.

1. Land-sea-ice features. In the Northern Hemisphere, the water features are the Arctic, the North Atlantic, and North Pacific Oceans. These bodies of water act as temperature moderators since they do not have large temperature variations. The major exception to this is when large areas are covered by ice in winter. The land features are the northern bulk of Eurasia, the smaller continent of North America, the island area of Greenland, and the Canadian Archipelago. As opposed to the water areas, the land areas tend to show the direct results of the extremes of seasonal heating and cooling by their seasonal temperature variations.

2. Mountains. The Arctic mountain ranges of Siberia and North America are factors which contribute to the climate and air-mass characteristics of the regions. These mountain barriers, as in midlatitudes, restrict the movement of air. During periods of weak circulation the air is blocked by the ranges and remains more or less stagnant over the area. It is during these periods that the air acquires the temperature and moisture characteristics of the underlying surface. Thus, these areas are air-mass source regions, and they are particularly effective as source regions during the winter when the surface is covered with snow and ice.

The Greenland icecap is essentially a mountain range and rises to a height in excess of 10,000 feet above mean sea level. It restricts the movement of weather systems, often causing low-pressure centers to move northward along the west coast of Greenland, resulting in some of the largest rates of falling pressure (other than hurricanes and tornadoes) recorded anywhere in the world. The deep low centers that move along the west coast of Greenland are, in the main, primarily responsible for the occasional high winds that are recorded in that area.

On some occasions, the winter temperatures in the Arctic are unusually high. This situation is brought about by deep low centers moving into the Arctic, coupled with compression of air (the foehn effect) as it often blows down off the sloping edges of the icecaps, primarily the Greenland icecap.

ARCTIC AIR MASSES.—The moisture content of air masses that originate over land is low at all altitudes in winter. The distinction between air masses almost disappears during the summer because of the nearly uniform surface conditions over the Arctic and subpolar regions. The frozen surface thaws under the influence of lengthened or continual daylight; the snow melts from the glaciers and pack ice; the ice melts in the lake areas in the Arctic; and the water areas of the polar basin increase markedly. Thus, the polar area becomes mild, humid, and semimaritime in character. Temperatures are usually between freezing and 50°F. Occasionally, strong disturbances from the south increase the temperature for short periods. Daily extremes, horizontal differences and day-to-day variabilities are slight.

During the winter months the air masses are formed over an area that is completely covered by ice and snow. The air masses are characterized by very cold surface air and a large temperature inversion in the lowest few thousand feet. Since the amount of moisture the air can hold is directly dependent on the temperature, the cold Arctic air is very dry (low absolute humidity). The air mass that originates over oceans does not have a surface temperature inversion in winter; the surface air temperature is warmer and there is a corresponding increase in the moisture content of the air. It is during movement inland of moist air from the warmer water that most of the rather infrequent Arctic cloudiness and precipitation occurs during this season.

During the summer months, the large expanse of open water and warmer temperatures result in a somewhat more abundant supply of moisture. Consequently, the largest amount of cloudiness and precipitation occurs during these summer months.

ARCTIC FRONTS.–The weather associated with a front in the Arctic has much the same cloud structure as with midlatitude fronts, except that the middle and high cloud types are generally much lower, and the precipitation is usually in the form of snow.

Periods of maximum surface wind usually occur during and just after a frontal passage. This strong windflow often creates hazards, such as blowing snow and turbulence which make operational flying difficult.

The best flying weather in the Arctic over land is most likely to occur in midsummer and midwinter; the worse (low ceilings and visibilities) in the transition period between the two seasons. Winter is characterized by frequent storms and well-defined frontal passages, but because of the dryness of the air, cloudiness and precipitation are at a minimum. In summer, there are fewer storm passages and fronts are weaker; however, the increased moisture in the air results in more widespread clouds and precipitation. Over the sea areas the summer weather is very foggy, but winds are of lower speeds than in winter.

During the transition periods of spring and fall, operational flying conditions are usually the worst. Frontal systems are usually well defined, active, and turbulent. Icing may extend to high levels.

TEMPERATURES IN THE ARCTIC.–Temperatures in the Arctic, as one might expect, are very cold most of the year. But contrary to common belief, the interior areas of Siberia, Northern Canada, and Alaska have pleasantly warm summers with many hours of sunshine each day. There are large differences in temperature between the interior and coastal areas.

In the interior during the summer days, temperatures often climb to the mid 60's or low 70's and frequently rise to the high 70's or low 80's, occasionally even into the 90's. Fort Yukon, Alaska, which is just north of the Arctic Circle has recorded an extreme high temperature of 100°F, while Verkhoyansk in north central Siberia has recorded 94°F.

During winter, the interior areas of Siberia, northern Canada, and Alaska act as a source region for the cold Arctic air that frequently moves southward into the middle latitudes. The coldest temperatures on record over the Northern Hemisphere have been established in Siberia. Oimekon, Siberia holds the record low with a temperature -108°F. Snag, in the Yukon Territory of Canada, witnessed the coldest temperature in North America with a record low of -83°F.

In the northern areas of the interior regions during the winter months, temperature are usually well below zero. In fact, during these long hours of darkness, the temperature normally falls to -20°F or -30°F, and in some isolated areas the normal daily minimum temperature may drop to -40°F. In north central Siberia the normal minimum daily temperature in the winter is between -45°F and -55°F.

The Arctic coastal regions, which include the Canadian Archipelago, are characterized by relatively cool, short summers. During the summer months the temperatures normally climb to the 40's or low 50's and occasionally reach the 60's. There is almost no growing season along the coasts, and the temperatures will fall below freezing during all months of the year. At Point Barrow, Alaska, the minimum temperature fails to fall below freezing on only about 42 days a year.

Over the Arctic Ocean the temperatures are very similar to those experienced along the coast; however, the summer temperatures are somewhat colder.

Winter temperatures along the Arctic coast are very cold but not nearly so cold as those observed in certain interior areas. Only on rare occasions does the temperature climb to above freezing during the winter months. The coldest readings for these coastal areas are in the -60's and -70's (degrees Fahrenheit).

These figures may seem surprising, since at first one might think that the temperatures near the North Pole would be colder than those over the northern continental interiors. Actually the flow of heat from the water under the ice has a moderating effect upon the temperature.

CLOUDINESS.–Cloudiness over the Arctic is at a minimum during the winter and spring and at a maximum during the summer and fall. The average number of cloudy days for the two 6-month periods on climatic charts shows a general decrease in cloudiness in the entire Arctic area during the winter months.

During the warm summer afternoons in the interior regions, scattered cumulus form and occasionally develop into thunderstorms. The thunderstorms are normally scattered and seldom form continuous lines. Along the Arctic coast and over the Arctic Ocean thunderstorms occur infrequently. Although tornadoes have been observed near the Arctic Circle, their occurrence is extremely rare.

WINDS.–Wind speeds are generally light in the continental interior during the entire year. The strongest winds in the interior normally occur during summer and fall. During winter, the interior continental regions are areas of strong anticyclonic activity which produce only light surface winds.

Strong winds occur more frequently along the Arctic coast than in the continental interiors. The frequency with which those high winds occur in coastal areas is greater in fall and winter than during summer. These winds frequently cause blowing snow.

Wind speeds greater than 70 knots have been observed at many Arctic coastal stations. Strong winds are infrequent over the icepack, but the wind blows continuously, since there is no hindrance offered by natural barriers such as hills and mountains. As a result of the combination of wind speed and low temperatures, the Arctic coastal area and the area over the icepack are very uncomfortable and limit outdoor human activity.

PRECIPITATION.–Precipitation amounts are small, varying from 5 to 15 inches annually in the continental interior and 3 to 7 inches along the Arctic coastal area and over the icepack. The climate over the Arctic Ocean and adjoining coastal areas is as dry as some of the desert regions in the United States. Most of the annual precipitation falls as snow on the Arctic Ocean and adjacent coastal areas and icecaps. On the other hand, most of the annual precipitation falls as rain over the interior.

RESTRICTION TO VISIBILITY.–Two conflicting factors make the subject of visibility in the polar regions very complex. Arctic air, being cold and dry, is exceptionally transparent and extreme ranges of visibility are possible. On the other hand, there is a lack of contrast between objects, particularly when all distinguishable objects are covered by a layer of new snow. Limitations to visibility in the Arctic are primarily blowing snow, fog, and local smoke. Local smoke is serious only in the vicinity of larger towns, and often occurs with shallow radiation fogs of winter.

1. Blowing snow. Blowing snow constitutes a more serious hazard to flying operations in the Arctic than in midlatitudes, because the snow is dry and fine, and is easily picked up by gentle and moderate winds. Winds in excess of 8 to 12 knots may raise the snow several feet off the ground, and the blowing snow may obscure surface objects, such as runway markers.

2. Fog. Of all the elements that restrict flying in the Arctic regions, fog in most respects is paramount. The two types of fog most frequently found in the polar regions are advection and radiation fog.

Fog is found most frequently along the coastal areas and usually lies in a belt parallel to the shore. In winter, the sea is warmer than the land. The moisture in the relatively warm air moving from the sea condenses over the cool land causing fog. This fog may be quite persistent.

3. Ice Fog. A fog condition peculiar to Arctic climates is ice fog. Ice fog is composed of minute ice crystals rather than water droplets of ordinary fog and is most likely to occur when the temperature is about -40°F or colder.

4. Sea Smoke or Steam Fog. The cold temperatures in Arctic may have effects which seem peculiar to people unfamiliar with the area. During the winter months, the inability of the air to hold moisture causes an unusual

phenomenon called sea smoke. This is caused by open bodies of comparatively warm water occurring simultaneously with low air temperatures. Actually, this phenomenon is about the same as the familiar one of steam forming over hot water.

In the case of sea smoke, the temperature of both the air and the water is much lower, but the air temperature is still the much colder of the two, causing steam to rise from the open water to form a fog layer. This fog occurs over open water, particularly over leads in the icepack, and is composed entirely of water droplets.

5. Arctic Haze. This is a condition of reduced horizontal and slant visibility (but good vertical visibility) encountered by aircraft in flight over Arctic regions. Color effects suggest this phenomenon to be caused by very small ice particles. Near the ground it is called Arctic mist or frost smoke; and when the sun shines on the ice particles, they are called diamond dust.

ARCTIC WEATHER PECULIARITIES.—The strong temperature inversions present over the Arctic during much of winter causes several interesting phenomena. Sound tends to carry great distances under these inversions. On some days, when the inversion is very strong, people's voices may be heard over extremely long distances as compared to the normal range of the human voice. Light rays are bent as they pass through the inversion at low angles. This may cause the appearance above the horizon of objects that are normally below the horizon. This effect, known as "looming," is a form of a mirage. Mirages of the type which distort the apparent shape of the sun, moon, or other objects near the horizon are common under inversion conditions.

One of the most interesting phenomena in the Arctic is aurora borealis (northern lights). These lights are by no means confined to the Arctic but are brightest at Arctic locations. Their intensity varies from a faint glow on certain nights to a glow which illuminates the surface of the earth with light almost equal to that of the light from a full moon.

The reactions resulting in the auroral glow have been observed to reach a maximum at an altitude of approximately 300,000 feet.

The amount of light from a snow-covered surface is much greater than the amount reflected from darker surfaces of the middle latitudes. As a result, useful illumination from equal sources is greater in the Arctic than in lower latitudes. When the sun is shining, sufficient light is often reflected from the snow surface to nearly obliterate shadows. This causes a lack of contrast, which in turn results in an inability to distinguish outlines of terrain or objects even at short distances. The landscape may merge into a featureless grayish-white field. Dark mountains in the distance may be easily recognized, but a crevasse (rift in a glacier or mass of land ice) immediately in front of one may be undetected due to lack of contrast. The situation is even worse when the unbroken snow cover is combined with a uniformly overcast sky and the light from the sky is about equal to that reflected from the snow cover. In this situation a person loses all sense of depth and orientation and appears to be engulfed in a uniformly white glow; the term for this optical phenomenon is whiteout.

Pilots have reported that the light from a half-moon over a snow-covered field is sufficient for landing purposes. It is possible on occasions to read a newspaper with the illumination from a full moon in the Arctic. Even the illumination from the stars creates visibility far beyond what one would expect elsewhere. It is only during periods of heavy overcast that the night darkness begins to approach the degree of darkness in lower latitudes. In lower latitudes, south of 65°N latitude, there will be long periods of moonlight. The moon may stay above the horizon for several days at a time.

Antarctic Weather

Many of the same peculiarities prevalent over the Arctic regions are also present in the Antarctic. For instance, the aurora borealis has its counterpart in the Southern Hemisphere. There the lights are called aurora australis. The same restrictions to visibility exist over the Antarctic regions as those over the Arctic. A few of the other characteristics of the Antarctic regions are as follows:

Precipitation occurs in all seasons, with the maximum probably occurring in summer. The amount of precipitation decreases poleward from the coast.

Temperatures are extremely cold. In winter, they decrease from the coast to the pole, but there is some doubt that this is true in summer. The annual variation of temperature as indicated by Little America shows the maximum in January and three separate minimums, with the lowest in early September. Other coastal stations show similar variations. A peculiar, and to date unexplained, feature of temperature variations during the Antarctic night is the occurrence of maximum temperature on cloudless days in the early hours after midnight. On cloudy days, the day is warmer than the night.

UNITED STATES WEATHER

The weather in the United States, with minor exceptions, is typical of all weather types within the temperate regions of the North American, European, and Asiatic continents. The general air circulation in the United States, as in the entire temperate zone of the Northern Hemisphere is from the west to the east. All closed surface weather systems (highs and lows) tend to move with this west-to-east circulation. However, since this is only the average circulation and weather systems move with the general flow, the fronts associated with the migratory lows move southward if they are cold fronts and northward if they are warm fronts. Surface low-pressure centers, with their associated weather and frontal systems, are often referred to as cyclones. Knowledge of the mean circulation in the temperate region makes it possible to observe and plot average storm tracks and to forecast future movement with a reasonable degree of accuracy.

Certain geographical and climatic conditions tend to make certain areas favorable for the development and formation of storm centers. In the United States these areas are west Texas, Cape Hatteras, central Idaho, and the northern portions of the Gulf of Mexico. Once a storm has formed, it generally follows the same mean track as the previous storm that formed in that area. The average or mean paths are referred to as storm tracks.

These storms are outbreaks on the polar front or the generation or regeneration of a storm along the trailing edge of an old front. The low pressure along these fronts intensifies in certain areas as the front surges southward ahead of a moving mass of cold air from the polar regions. Much of the weather in the Temperate Zone is a direct result of the effect of these storms (especially in winter). In addition to the weather from the effect of these storms is air-mass weather. Air-mass weather is the name given to all weather other than the frontal weather in the temperate region. Air-mass weather is the net effect of local surface circulation, terrain, and the modifying effect of significant water bodies. This effect is discussed more in detail in chapter 6.

There are many subdivisions of weather regions in the United States. For the purpose of this discussion, we have divided the United States into the regions as indicated in figure 2-10.

Northwest Pacific Coast Area

The northwest Pacific coast area has more precipitation than any other region in North America. Its weather is the result of frontal phenomena, consisting mainly of occlusions which move in over the coast from the area of the Aleutian low to the coast, and orographic lifting of moist stable maritime air. Predominant cloud forms are stratus and fog which are common to all seasons. Rainfall is most frequent in winter and least frequent in summer.

Southwest Pacific Coast Area

This region experiences a Mediterranean type climate and is distinctively different from any other North American climate. This type climate occurs exclusively in the Mediterranean and Southern California in the Northern Hemisphere; in the Southern Hemisphere it occurs over small areas of Chile, South Africa, and Southern Australia.

This type climate is characterized by warm to hot summers, tempered by sea breezes, and by mild winters during which the temperatures seldom go below freezing. Little or no rainfall occurs in the summer and only light to moderate rain in the winter.

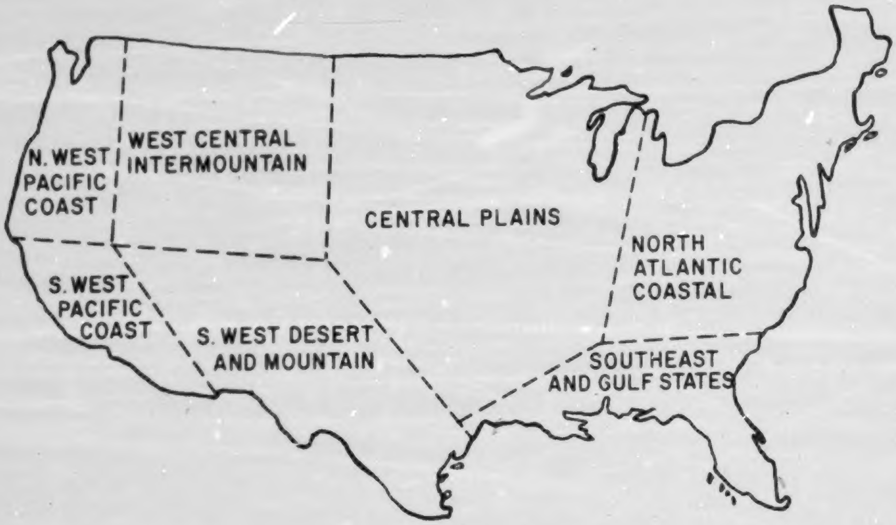
N. WEST PACIFIC COAST
WEST CENTRAL INTERMOUNTAIN
CENTRAL PLAINS
NORTH ATLANTIC COASTAL
S. WEST PACIFIC COAST
S. WEST DESERT AND MOUNTAIN
SOUTHEAST AND GULF STATES

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Figure 2-10.—United States weather regions.

Cold fronts rarely penetrate the southwest Pacific coast region. The weather over this region is due to the circulation of moist Pacific air from the west being forced up the slope of the coastal range. In summer, air is stable, and stratus and fog result. In winter, unstable air, which is forced over the mountain ranges, causes showers and sometimes snow showers in the mountains.

Intermountain West Central Area

This area includes the Great Plains region. This region is located east of the Cascade and Coastal Ranges and west of the Mississippi Valley, and north of the southwest desert area. The climate is generally cold and dry in the winter, and warm and dry in summer. Most of the region is semiarid. The western mountain range, which acts as a climatic barrier, has an extreme drying effect on the air in the westerly circulation.

The maximum rainfall occurs in the spring and is due mainly to the predominance of cyclonic storm track passages during this season. In midwinter a cold high is generally centered in this region and generally prevents the possibility of storm passages. Annual precipitation is light.

Southwest Desert and Mountain Area

This area includes Lower California and some of Southeast California and all the southern half of Arizona and New Mexico and west and southwest Texas. It is an area almost completely surrounded by high mountains and is either all very arid or actual desert. Annual rainfall seldom exceeds 5 inches. The more northerly sections have cold winters, and all parts have extremely hot summers. The chief flying hazard is the predominance of summer and spring thunderstorms caused mainly by maritime tropical air penetrating into this area with the expansion of the Bermuda high in summer and the air being forced aloft at the mountains. For this reason nearly all significant peaks and ranges have thundershowers building over them in spring and summer. The thunderstorms are generally scattered but always severe; however, they may generally be avoided by circumnavigation.

Central Plains Area

This area includes the continental climate regions of the Great Plains, Mississippi Valley, and Appalachian Plateau between the Rocky Mountains to the west and the Appalachians to the east and the Gulf States on the south.

In the western section it is rather dry but moisture increases eastward. The main weather hazards are caused by wintertime outbreaks along the polar front and associated wave phenomena. Also, the development of convective type air-mass thunderstorms which are prevalent over this area in summer.

Frontal passages, both cold and warm, with associated weather are common in this area. Thunderstorms are usually of convective origin and are most violent if they have developed in maritime tropical air. This occurs often in the spring, and tornado activity becomes a climatic feature due to its frequency.

Southeast and Gulf States Area

This area includes all the states bordering on the Gulf of Mexico and South Carolina and Georgia. A circulation phenomenon known as gulf stratus affects this area. The stagnation of southward moving cold fronts, rapidly moving squall lines, air-mass thunderstorms, and the gulf stratus occurring in various combinations make this area an especially difficult one for the forecaster.

Frontal passages may be expected only in the late fall, winter, and early spring. In the winter, when the circulation near the surface is southerly, the warm moist gulf air is cooled from below to saturation. When this occurs, fog and stratus may form and persist over the area for several days. The southerly circulation in summer causes warm moist air to be heated from below, and convective thunderstorms are common. Since the air is generally quite moist and unstable, the storms are generally severe.

North Atlantic Coastal Area

This is an area of storm track convergence, and cyclonic storm activity with accompanying weather is frequent in winter. Moreover, this is an area in which these storms are intensified by heating and addition of moisture over the Great Lakes. The lake effect is directly accountable for the great amounts of snowfall often found over this area in the winter. Generally good weather prevails in summer due to the dominating influence of the Bermuda high.

CHAPTER 3

ATMOSPHERIC PHYSICS

The science of physics is devoted to finding, defining, and reaching the solutions to problems. Physics, therefore, not only breeds curiosity of the surrounding environment, but it provides a means of acquiring answers to questions which continue to arise.

The basic physical subjects relating to motion, force, and energy as they pertain to meteorology are briefly presented in chapter 3 of Aerographer's Mate 3 & 2, NavTra 10363-D. It is recommended that this material be reviewed prior to continuing with the more complicated processes to be presented in this manual.

VECTORIAL MOTION

FORCE

Many people think that all force comes from muscular effort, such as the effort required to push a box resting on the deck. However, water in a can exerts force on the sides and bottom of the can and an upward force on any object on the surface of the water. A tug exerts force on the ship it is pushing or pulling. In all of these examples, the body exerting the force is in contact with the body on which the force is exerted; forces of this type are called contact forces. There are also forces which act through empty space without contact. The force of gravity exerted on a body by the earth—known as the weight of the body—is one example of this type of force. Forces which act through empty space without contact are called action-at-a-distance forces. Electric and magnetic forces are action-at-a-distance forces.

Vectors

Problems arise in which it is necessary to deal with one or more forces acting on a body. To solve problems involving forces, a means of representing forces must be found. A force is completely described when its magnitude, direction, and point of application are given. Because a vector is a line which represents both magnitude and direction, a vector may be used to describe a force. The length of the line represents the magnitude of the force, the direction of the line represents the direction in which the force is being applied, and the starting point of the line represents the point of application of the force. For example, to represent a force of 10 pounds acting due east on point A, draw a line 10 units long, starting at point A and extending in a direction 90° clockwise from north. (See fig. 3-1.)

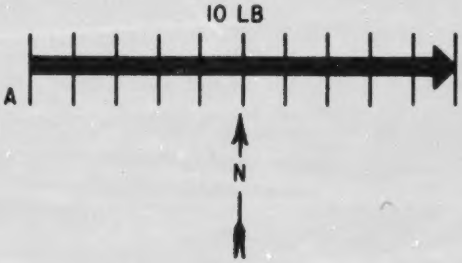

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Figure 3-1.—Example of a vector.

Composition of Forces

If two or more forces are acting simultaneously at a point, the same effect can be produced by a single force of the proper size and direction. This single force, which is equivalent to the action of two or more forces, is called the resultant. Putting component forces together to find the resultant force is called composition of forces. The vectors representing the forces must be added to find the resultant. Because a vector represents both magnitude and direction, the method for adding vectors differs from the procedure used for scalar quantities, which have no direction.

Consider this example: Find the resultant force when a vertical force of 5 pounds and a horizontal force of 10 pounds are applied at point A. (See fig. 3-2.)

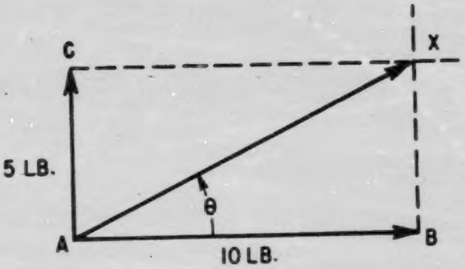

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Figure 3-2.—Composition of two right-angle forces.

The resultant force may be found as follows: Represent the given forces by vectors AB and AC drawn to a suitable scale, as in figure 3-2. At points B and C draw dashed lines perpendicular to AB and AC, respectively. From A draw a line to the point of intersection X of the dashed lines. Vector AX represents the resultant of the two forces. Thus, when two mutually perpendicular forces act on a point, the vector representing the resultant force is the diagonal of a rectangle. The length of AX, on the same scale as that for the two original forces, is the size of the resultant force; the angle θ gives the direction with respect to the horizontal.

When it is desired to find the resultant of two forces which are not at right angles, the following graphic method may be used:

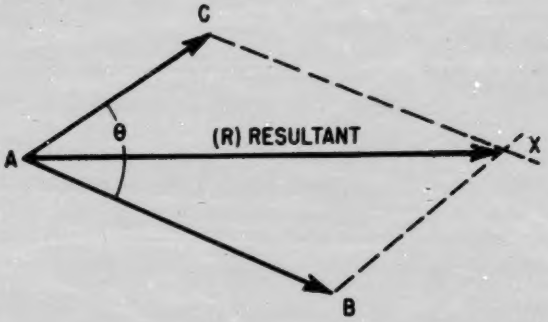

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Figure 3-3.—Graphic method of the composition of forces.

Let AB and AC (fig. 3-3) represent the two forces drawn accurately to scale. From point C draw a line parallel to AB and from point B draw a line parallel to AC. The lines will intersect at a point X, as shown in figure 3-3. The force AX is the resultant of the two forces AC and AB. Note that the two dashed lines and the two given forces make a parallelogram ACXB. Solving for the resultant in this manner is called the parallelogram method. The size and direction of the resultant may be found by measurement from the figure drawn to scale. This method applies to any two forces acting on a point whether they act at right angles or not. Note that the parallelogram becomes a rectangle for forces acting at right angles.

Consider the following problems as practical examples of combining forces:

1. A supply barge is anchored in a river during a storm. If the wind acts westward on it with a force of 4,000 pounds and the tide acts southward on it with a force of 3,000 pounds, what is the direction and the magnitude of the resultant force acting upon the barge? Figure 3-4 shows the forces acting upon the barge.

If the force vectors have been drawn to scale, the magnitude and direction of the resultant may be obtained by measuring the length of the diagonal of the completed parallelogram and the angle the diagonal makes with a side of the parallelogram. The resultant can be found also by geometry and trigonometry.

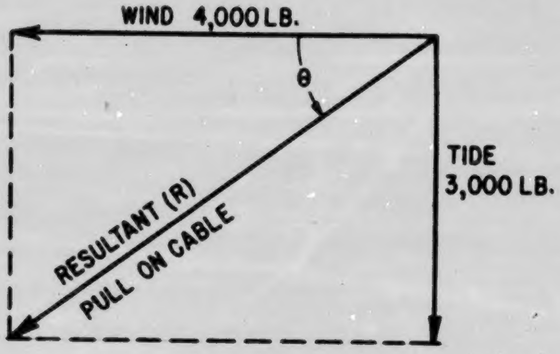

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Figure 3-4.—Problem using the composition of forces.

$$R^2 = 3000^2 + 4000^2$$

$$R^2 = (3 \times 1{,}000)^2 + (4 \times 1{,}000)^2$$

$$R^2 = (3^2 \times 1{,}000^2) + (4^2 \times 1{,}000^2)$$

$$R^2 = (1{,}000^2) \times (3^2 + 4^2)$$

$$R^2 = (1{,}000^2) \times (25)$$

$$R = \sqrt{(1{,}000^2) \times (25)} = (1{,}000) \times (5)$$

$$R = 5{,}000 \text{ lb}$$

From trigonometry:

$$\tan \theta = \frac{\text{opposite side}}{\text{adjacent side}}$$

$$\tan \theta = \frac{3{,}000}{4{,}000} = 0.750$$

The angle θ can be found from $\tan \theta$ by using either trigonometric tables or the slide rule.

$$\theta = 36.9^\circ$$

The resultant is a force of 5,000 pounds acting at an angle of 36.9° with the direction of the wind or in a direction of 233.1° (270° - 36.9° = 233.1°).

2. With a slight modification, the parallelogram method of addition applies also to the reverse operation of subtraction. Consider the problem of subtracting force AC from force AB. (See fig. 3-5.)

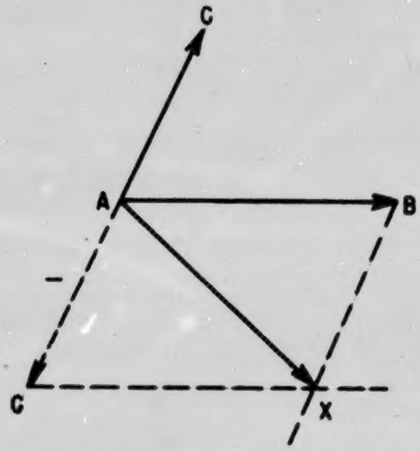

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Figure 3-5.—Parallelogram method of subtracting forces.

First, force AC is reversed in direction giving -AC. Then, forces -AC and AB are added by the parallelogram method, giving the resulting AX, which in this case is the difference between forces AB and AC. A simple check to verify the results consists of adding AX to AC; the sum or resultant should be identical with AB.

MOTION

Everything about us in the world moves. Even a body supposedly at rest on the surface of the earth is in motion because it is actually moving with the rotation of the earth and the earth is turning in its orbit around the sun. Therefore, the terms "rest" and "motion" are relative terms. The change in position of any portion of matter is motion. The atmosphere, being a gas, is subject to much motion. Temperature, pressure, and density act to produce the motions of the atmosphere. The motions are subject to well-defined physical laws.

Newton's Laws of Motion

Newton's first law of motion deals with the principle of inertia and states that every body continues in a state of rest or of uniform motion in a straight line unless acted upon by some external force.

The second law states that whenever a net force acts on a body, it produces an acceleration in the direction of the net force. The acceleration is directly proportional to the net force and inversely proportional to the mass of the body.

Newton's third law of motion concerns the principle of action and reaction, and states that to every action there is an equal and opposite reaction.

GAS LAWS

In order for you as Aerographer's Mates to have a general understanding of the earth's atmosphere, it is necessary for you to have a working knowledge of the gas laws. From these you can readily discern that the behavior of any gas depends upon the variations in temperature, pressure, and density. Since the atmosphere is a mixture of gases, its behavior is governed by well-defined laws. Although these gas laws were discussed in AG 3&2, NT 10363-D, they are presented briefly in the following paragraphs to provide continuity prior to discussing their application.

A real and ideal gas is a gas which obeys Charles' and Boyle's laws perfectly. The atmosphere very nearly obeys these laws, and meteorologists consider it as a perfect gas in most cases.

BOYLE'S LAW states that if the temperature remains constant, the volume of an enclosed gas is inversely proportional to its pressure. The formula for Boyle's law is as follows:

$$VP = V'P'$$

V – initial volume
P – initial pressure
V' – new volume
P' – new pressure

CHARLES' LAW states that if the pressure remains constant, the volume of a gas is directly proportional to its temperature. Experiments show that the volume will increase by 1/273 for a 1°C rise in temperature. The formula for Charles' Law is as follows:

$$VT' = V'T, \text{ or it can be written } \frac{V}{V'} = \frac{T}{T'}$$

V – initial volume
T – initial temperature (absolute)
V' – new volume
T' – new temperature (absolute)

The UNIVERSAL GAS LAW is a combination of Boyle's law and Charles' law. It states that the product of the pressure and volume of a gas is directly proportional to the absolute temperature. The formula is as follows:

$$PVT' = P'V'T$$

P – initial pressure
V – initial volume
T – initial temperature (absolute)
P' – new pressure
V' – new volume
T' – new temperature (absolute)

EQUATION OF STATE

The EQUATION OF STATE is a general formula which gives the same information as Boyle's law and Charles' law. It involves a gas constant, which is a value assigned each gas. For instance, the gas constant of air is 2,870 when the pressure is expressed in millibars, and the density is expressed in metric tons per cubic meter or in grams per cubic centimeter. The constant may be expressed differently, depending on the system of units used. The following formula is an expression of the equation:

$$P = \rho RT$$

P – pressure in millibars
ρ – density
R – gas constant
T – temperature (absolute)

DALTON'S GAS LAWS

John Dalton, an English physicist, formulated two fundamental gas laws relative to the pressure of a mixture of gases. One of the laws states that a mixture of several gases which do not react chemically exerts a pressure equal to the sum of the pressures which the several gases would exert separately if each were allowed to occupy the entire space alone at the given temperature. The other law states that the total pressure exerted by the mixture of gases is equal to the sum of the partial pressures.

Water vapor in the atmosphere, for instance is independent of the presence of other gases; the vapor pressure is independent of the pressure of the dry gases in the atmosphere. The vapor pressure for any given temperature has a maximum limit, reached when the air is saturated. The total atmospheric pressure is found by adding the pressures of the dry air and the water vapor.

AVOGADRO'S NUMBER

It was the hypothesis of Avogadro that equal volumes of all gases under the same pressure and temperature contain the same number of molecules. The number of molecules in a gram molecule of gas is known as AVOGADRO'S NUMBER: a gram molecule of any gas contains 6.02×10^{23} molecules and at 0°C and 760 mm pressure occupies a volume of 22,421 cm^3. A gram molecule is the mass of a compound equal numerically to the value of its molecular weight; likewise, a gram atom is the mass of an element numerically equal to the value of its atomic weight.

STANDARD CONDITIONS FOR GASES

The conditions under which gases must be compared, densities determined, and gas constants derived, are known as the STANDARD CONDITIONS FOR GASES. The standard conditions are a pressure of 760 mm of mercury and a temperature of 0°C.

HYDROSTATIC EQUATION

The HYDROSTATIC LAW states that the difference in pressure between two points at different levels in a mass of fluid at rest is equal to the weight of a column of the fluid of a unit cross section reaching vertically from one level to another.

For the atmosphere, the hydrostatic law states that the difference in pressure between two points in the atmosphere, one above the other, is equal to the weight of the air column between the two points. Although these two laws are essentially the same, there are two variables which must be considered when applied to the atmosphere. They are temperature and density.

From Charles' law it was learned when the temperature increases, the volume increases and the density decreases. Therefore, the thickness of a layer of air will be greater when the temperature is increased. To find the height of a pressure surface in the atmosphere (such as in working up an adiabatic chart) these two variables must be taken into consideration.

In the computation of a radiosonde observation, a set of tables has been computed and the density has been incorporated in these tables.

The thickness of a layer can be determined by the following formula:

$$Z = (49{,}080 + 107t)\frac{Po - P}{Po + P}$$

Z – altitude difference in feet (thickness of layer)
t – mean temperature in degrees Fahrenheit
Po – pressure at the bottom point of the layer
P – pressure at the top point of the layer

For example, let us assume that a layer of air between 800 and 700 millibars has a mean temperature of 30°F. Applying the formula, the following value is obtained:

$$Z = (49{,}080 + 107 \times 30)\frac{800 - 700}{800 + 700}$$

$$Z = (49{,}080 + 3{,}210)\,\frac{100}{1{,}500}$$

$$Z = (52{,}290)\,\frac{1}{15}$$

Z = 3,486 feet (1,063 meters)
(1 meter = 3.28 feet)

APPLICATION OF THE GAS LAWS

In the field of aviation, and particularly in the Naval Weather Service, it is necessary to be able to apply the gas laws to many situations. For example, in the use of weather balloons we mix helium with a small quantity of air. As we add the gases to the balloons, the pressure increases on the walls, but at the same time it is becoming lighter and lighter until it floats. Since helium is lighter than air, the mixture has become lighter and less dense. This is a direct application of Dalton's law.

The hydrostatic law states that the difference in pressure between two points is equal to the weight of the air column between these two points. We saw in our application of this law that the thickness of the column in question is dependent upon the mean virtual temperature of the column. Virtual temperature is defined as the temperature the air would have if all water vapor were removed from it. It is expressed by the following formula:

$$T_v = \frac{T}{1 - (0.379\frac{e}{p})}$$

where

T_v is the virtual temperature
T is the temperature in degrees Celsius
0.379 is a constant
e is the vapor pressure
p is the atmospheric pressure

When the thickness is decreased, the mean virtual temperature decreases, and the density increases in the air column. This illustrates Charles' law. For layers of equal pressure difference, the thickness of the layers increase as we go up into the atmosphere due to decreased pressure on the layers, and consequently a decrease in the density of the strata. This is an illustration of Boyle's law.

Since there is a definite relationship of thickness to temperature, this relationship is put to work in a system called differential analysis. The height or difference in thickness between two isobaric surfaces is found by applying the formula or by subtracting the height of the lower level from the height of the upper level. These lines of equal thickness are considered and could be labeled mean isotherms for these layers, showing advection of cold or warm air.

ENERGY CONSIDERATIONS–STABILITY AND INSTABILITY

ATMOSPHERIC ENERGY CONSIDERATIONS

Energy has previously been defined as the capacity to perform work. There are two basic kinds of energy. They are KINETIC and POTENTIAL. Kinetic energy is ability to perform work due to motion. Potential energy is ability to perform work due to position or condition. The term "kinetic" is used for energy due to present motion, whereas the term "potential" applies to energy stored for later action.

The kinetic theory of gases is very helpful in understanding the behavior of gases. Gases, like other substances, consist of molecules. Unlike solids, molecules of gas have no inherent tendency to stay in one place. Instead, gas molecules, since they are smaller than the space between them, move about at random (but in straight lines until they collide with each other or with other obstructions). Their movement has an average speed at a given temperature. When gas is enclosed, its pressure depends on the number of times the molecules strike the surrounding walls in a unit of time. The number of times the molecules strike per unit of time against the walls remains constant as long as the temperature and the volume remain constant. If the volume (the space occupied by the gas) is decreased, the density of the gas is increased, and the number of blows against the walls is increased, thereby increasing the pressure. When the temperature is increased, there is a corresponding increase in the speed of the molecules; they strike the walls at a faster rate, increasing the pressure, provided that the volume remains constant.

According to the kinetic theory of gases, the temperature of a gas is dependent upon the rate at which the molecules are moving about and is proportional to the kinetic energy of the moving molecules. The kinetic energy of the moving molecules of a gas is the internal energy of the gas, and it follows that an increase in temperature is accompanied by an increase in the internal energy of the gas. Likewise, an increase in the internal energy results in an increase in the temperature of the gas.

An increase in the temperature of a gas or in its internal energy can be produced by the addition of heat or by performing work on the gas. A combination of these can likewise produce an increase in temperature or internal energy. This is in accordance with the first law of thermodynamics.

In the application of the first law of thermodynamics to a gas, it may be said that the two main forms of energy are internal energy and work energy. Internal energy is manifested as sensible heat or temperature; work energy is manifested as pressure changes in the gas. In other words, work is required to increase the pressure of a gas and work is done by the gas when the pressure diminishes. It follows, then that if internal (heat) energy is added to a simple gas this energy must show up as an increase in either temperature or pressure, or both. Also, if work is performed on the gas, the work energy must show up as an increase in either pressure or temperature, or both.

Consider air in a cylinder, which is enclosed by a piston. In accordance with the first law of thermodynamics, any increase in the pressure exerted by the piston results in work being done on the air. As a consequence, either the temperature and pressure must be increased or the heat equivalent of this work must be transmitted to the surrounding bodies. In the case of a plain compressor, this work done by a piston is changed into an increase in the temperature and the pressure of the air. It also results in some increase in the temperature of the surrounding body.

If the surrounding body is considered to be insulated so that it is not heated, there is no heat transferred, and the air must acquire this additional energy as an increase in temperature and pressure. The process by which a gas, such as air, is heated or cooled without heat being added to the gas or taken away is called an adiabatic process.

In the atmosphere, adiabatic and nonadiabatic processes are taking place continuously. The air near the ground is receiving heat from or giving heat to the ground. These are nonadiabatic processes. However, in the free atmosphere somewhat removed from the earth's surface, the short-period processes are adiabatic. When a parcel of air is lifted in the free atmosphere, it encounters areas of decreasing pressure. To equalize this pressure, the parcel must expand. In expanding, it is doing work. In doing work, it uses heat. This results in a lowering of temperature, as well as a decrease in the pressure and density. When a parcel of air descends in the free atmosphere, it encounters areas of increasing pressure. To equalize the pressure, the parcel must contract. In doing this, work is done on the parcel. This work energy, which is being added to the parcel, shows up as an increase in temperature. The pressure and density increase in this case, too.

LAPSE RATES

The rate of cooling that a parcel of air undergoes as it ascends in the free atmosphere (or the rate of heating as it descends) is known as the ADIABATIC LAPSE RATE. For unsaturated air the rate of change is 1°C per 100 meters or 5 1/2°F per 1,000 feet. This is known as the DRY ADIABATIC LAPSE RATE. For saturated air the rate of change is different. When a parcel of saturated air ascends in the free atmosphere, the rate of change is known as the SATURATION ADIABATIC LAPSE RATE.

The saturation adiabatic lapse rate is the result of the condensation that takes place in a saturated parcel of air as it ascends above the condensation level. For each gram of water condensed, about 600 calories of heat are liberated. The latent heat of condensation is absorbed by the air. Consequently, the lapse rate becomes less than the dry adiabatic lapse rate.

The mean slope of the saturation adiabat may be taken as approximately 0.55°C per 100 meters or 3°F per 1,000 feet. The term MEAN SLOPE is used because the saturation adiabatic lapse rate increases with altitude. This is a result of the decrease of water vapor with altitude, consequently a decrease in the total heat of condensation which is liberated.

The normal or average decrease of temperature with height is known as the normal or average lapse rate. The normal lapse rate is about 3 1/2°F per 1,000 feet up to the tropopause. However, the actual lapse rate in the atmosphere at any given time depends on turbulence, radiation processes, conduction of heat near the ground, or the transport of air by horizontal advection in the upper layers. Condensation of moisture or evaporation also affect the lapse rate by the addition or removal of the heat of condensation. Figure 3-6 shows some of the various types of lapse rate which may be found in the atmosphere.

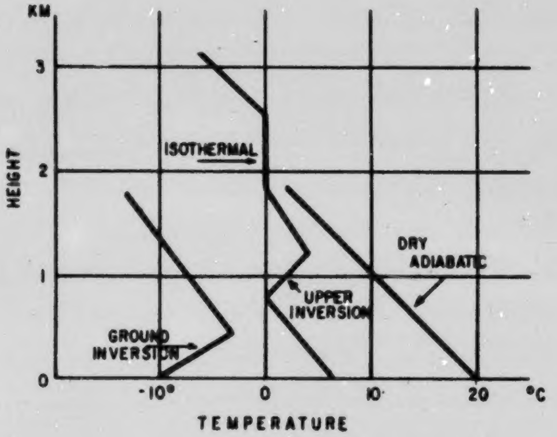

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Figure 3-6.—Lapse rates in the atmosphere.

Reversible Process

The REVERSIBLE PROCESS is based upon the assumption that no condensed water falls as precipitation, but is carried with the ascending parcel. The ascending parcel undergoes several stages as follows:

1. The dry stage, where the parcel is lifted dry adiabatically to saturation.
2. The rain stage, where all water vapor exceeding the saturation amount is condensed to liquid water.
3. The hail stage, occurring at 0°C when the parcel rises isothermally because the expansional cooling is offset by the heat of fusion being released when the liquid water is frozen.
4. The snow stage, when the excess moisture is changed directly from a vapor to a solid (snow).

The process is reversible, since the parcel reaches the top of the atmosphere with the same water content with which it started. Upon its descent, the same stages will occur in reverse order and the parcel arrives at the original level (pressure) with the same temperature as before.

Irreversible Process

The IRREVERSIBLE PROCESS is based upon the assumption that all excess water which is condensed from the air will fall immediately as precipitation. The stages in this process are nearly similar to the reversible process.

1. The dry stage is the same as the reversible.
2. The rain stage is the same as the reversible except that moisture falls as rain.
3. The hail stage is eliminated, since there is no liquid to change into ice.
4. The snow stage is the same as the reversible except that excess moisture condenses and falls as snow.

The parcel is dry by the time it reaches the top of the atmosphere; therefore, it must descend dry adiabatically, and the temperature is therefore much higher upon reaching the original level.

The irreversible process closely approximates the actual conditions in the atmosphere; this is reflected on the Skew-T diagram by the saturation adiabats showing no hail stage.

Figure 3-7 shows the various stages during the ascent and descent of air in the irreversible process.

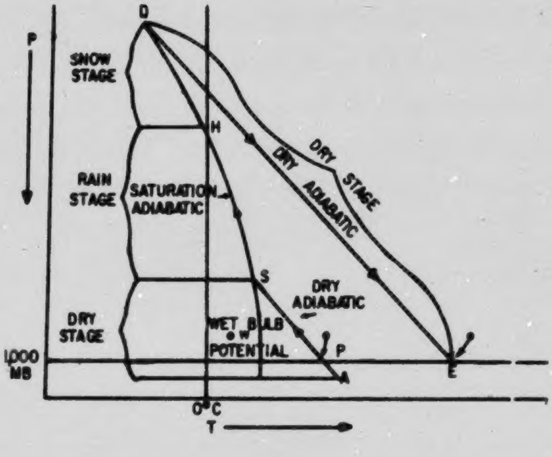

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Figure 3-7.—The irreversible process.

If a parcel of air (A) follows the dry adiabatic lapse rate to the 1,000-mb line, the temperature at that point (P) is the potential temperature. If the parcel continues up the dry adiabat to the saturation point (S), condensation will begin and the parcel will follow the saturation adiabat through the rain stage (S-H). When the temperature falls below freezing, the condensation will be in the form of snow (H-D). Actually in nature supercooled water droplets may form even below the freezing point. When all the moisture has condensed, all of the latent heat of condensation has been added to the air, and if the parcel is then brought back to the 1,000-mb level, it will follow the dry adiabatic lapse rate (D-E) and arrive back at this level with a temperature of E. This hypothetical temperature is called the equivalent potential temperature.

CONDENSATION AND PRECIPITATION

The CLASSIC CONDENSATION THEORY is the theory used when making thermodynamic computations, such as on the Skew T diagram. It is perfectly valid for these computations. In this theory it is assumed that water is entirely in vapor form until 100 percent relative humidity is reached, and then it changes to liquid or ice. It assumes that liquid drops do not exist at a temperature below freezing; and below freezing only direct crystallization or sublimation occurs.

When attempting to explain the actual process of condensation and precipitation in the atmosphere, the CLASSIC CONDENSATION THEORY is no longer completely valid. These are some of the defects in the theory:

1. Clouds, and especially fog, are likely to occur at less than 100 percent relative humidity. The whole process of formation of a droplet is a continuous one that, however, is most rapid at 100 percent humidity.

2. Liquid droplets supercooled to temperatures several degrees below freezing are so common in the atmosphere as to be regarded as the rule rather than the exception.

3. Liquid drops not only exist at temperatures below freezing, but new condensation occurs at these temperatures as well as direct sublimation.

Before condensation can occur in the free atmosphere, the temperature of the air must be reduced to near the dewpoint, or the moisture content must be increased so as to increase the dewpoint to near the temperature. In laboratory experiments, it has been proved that even these conditions will not induce condensation if the air is pure. It was proved that in pure air a supersaturation of 400 percent was possible before condensation occurred.

There are also several other observations which do not conform to these theories. Drizzle may fall out of stratus or stratocumulus layers that do not extend into the freezing temperatures, particularly at sea and in coastal districts with onshore winds. In the Tropics, and also in warm maritime air masses in temperate latitudes, cumulus has been observed to yield light, moderate, and even heavy rain though the tops did not extend above the freezing level.

The explanation for the latter phenomenon appears to be that in the Tropics, where the freezing level is high, cumulus can develop to such great depths without reaching 0°C, that coalescence between cloud droplets becomes effective enough to result in appreciable rain. When the cloud droplets are of initially different sizes and hence have different settling rates, some of them will collide and coalesce. This increases their settling rate, and consequently, the number of collisions per unit of time. By such a chain reaction, the droplets grow to sufficient size to fall out as precipitation. A cloud of great vertical extent is needed for this process to result in heavy precipitation. This condition is satisfied in the Tropics. The high temperatures found there give a high liquid-water content, which also furthers the coalescence process.

The drizzle which falls out of nonfreezing layer clouds is a more gentle display of the coalescence process. Drizzle appears to be more frequent over the sea and along the sea borders, other things being equal; this fact favors the part played by salt nuclei. Precipitation from nonfreezing clouds has not been noted in continental air masses.

From these observations we may reach the following conclusions. The coalescence process may account for much of the precipitation which falls in the Tropics; the classic theory, on the other hand, applies to most of the precipitation occuring in middle and high latitudes.

Whenever moderate or heavy rain falls in the temperate or Arctic regions, it originates mainly in clouds that, in the upper portions at least, have reached negative Celsius temperatures.

A further conclusion based on many studies of temperature of the cloud top as a condition for precipitation reveals that when rain or snow–continuous or intermittent–reaches the ground from stratiform clouds, the clouds–solid or layered–extend in most cases to heights where temperatures are below -12°C or even -20°C.

Nuclei

The foreign particles in the air may be divided into three classes:

1. Hygroscopic nuclei.
2. Sublimation nuclei.
3. Neutral or nonhygroscopic particles.

When air is cooled so that it approaches the dewpoint, the hygroscopic nuclei begin to absorb water from the air. The larger nuclei will then cause condensation to occur, even before the saturation point is reached. However, as the drops grow in size they become so diluted that they become less and less active as hygroscopic material. The condensation process can then proceed only when the air is cooled slightly below its dewpoint so that a slight amount of supersaturation is present. It can be seen that condensation on hygroscopic nuclei is a continuous process, beginning at low relative humidities.

The most common hygroscopic nuclei are sea salts, sulfuric acids, and nitric acids. Hygroscopic nuclei vary in number, with the maximum over cities, decreasing in number over rural areas, and at a minimum over the oceans. Annual variations in amounts of hygroscopic nuclei are caused by the increased amount of combustion taking place during the winter months. Diurnal variation is caused by the lack of sunlight at night to oxidize sulfur and nitric dioxides. Since the source region for hygroscopic nuclei is at the surface of the earth, they decrease in number with altitude.

Sublimation nuclei are much smaller and fewer in number than hygroscopic nuclei. They are shaped like an ice crystal. Ice forms on these nuclei below water saturation, but at or above ice saturation. Sublimation nuclei are not very active between 0°C and -10°C; in fact, when considering the entire atmosphere at all temperatures down to about -40°C, there are more liquid droplets than ice particles.

Neutral or nonhygroscopic particles are particles such as ordinary dust. These particles may act as condensation nuclei, but seldom do.

The atmosphere has been described as an AEROSOL, which is a colloidal system in which the dispersed water vapor is composed of either solid or liquid particles, and in which the dispersing medium is the air. A cloud has been described as colloidally unstable by virtue of its position in a turbulent atmosphere.

There have been many theories presented on the processes leading to colloidal instability within a cloud which would cause the growth of raindrops. A few of the more feasible ones are as follows:

1. Electrical attraction.
2. Hydrodynamical attraction.
3. Vapor pressure gradient from smaller to larger drops.
4. Introduction of extremely hygroscopic nuclei.
5. Collision due to turbulence.
6. Vapor pressure gradient from warmer to colder drops.
7. Vapor pressure gradient from liquid to ice (Bergeron-Findeisen Theory).
8. Nonuniform drops in the gravitational field.

The last two theories are considered to be the most important in the formation of raindrops. The liquid-to-ice theory is the most important during the initial formation of the droplet, but once they have grown to such size that they begin to fall, the gravitational field theory becomes the most important.

Cloud and Weather Modification Methods

Activity in cloud and weather modification has been on a sound and realistic basis only since 1958.

Cloud modification methods have been mainly in the use of dry ice, silver iodide, and water to increase the cloud amount and to possibly trigger precipitation.

The object of seeding with dry ice is to cause the coexistence of ice and water. Seeding with dry ice may be used to dissipate clouds, to precipitate clouds, or to make existing clouds more persistent.

The use of silver iodide has been found to be the most effective source of ice crystal nuclei found to date and is most effective at temperatures below - 10°C.

The use of water attempts to employ the principle of nonuniform drops in the gravitational field. It may be used on actively convective portions of large cumulus clouds.

According to Schaefer, the most favorable atmospheric conditions for precipitating clouds by seeding are large cumulus clouds without precipitation already occurring, abundant moisture, a large lapse rate, a low concentration of ice nuclei, the absence of wind shear, and either no inversions or few inversions, or weak ones.

In practical application of the weather modification program the prevention of thunderstorm formation, hail, windstorms, and torrential rain may be accomplished by either dissipating or overseeding, or increasing precipitation by seeding the clouds with the proper amounts. No evidence appears to exist that clouds can be milked of their moisture in flat regions, but indications are that seeding, especially with silver iodide smoke, can increase precipitation as much as 10 to 15 percent.

Another application of this process is the so-called CLOUDBUSTERS of the Air Force. They have found that holes or windows can be punched in certain clouds which hinder aircraft landings and takeoffs, parachute drops, and rescue and reconnaissance missions. Windows more than 3 miles wide have been created by overseeding such clouds with dry ice pellets.

The dissipation of certain types of fogs can be accomplished by seeding. This is more effective for cold fogs and has little effect on warm or ice crystal fogs.

STABILITY AND INSTABILITY

Most weather phenomena depend upon whether the air masses are stable or unstable. As stated before, moisture content plays an important part in weather. A parcel of air may be stable when dry and then may become unstable if it is lifted to saturation. An understanding of stability and instability is therefore essential to a forecaster.

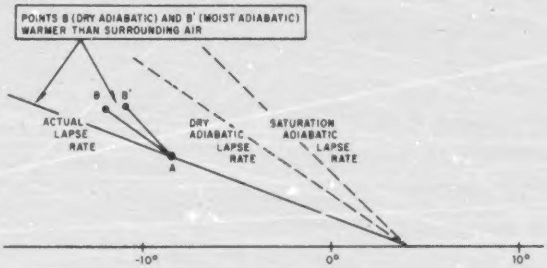
STABILITY is the state of equilibrium in which a parcel of air has a tendency to resist displacement from the level at which it is in equilibrium with its environment or, if displaced, to return to its original position. INSTABILITY is the state of equilibrium in which a parcel of air when displaced has a tendency to move farther away from its original position.

The stability and instability of air depend a great deal on the moisture content of the air. Therefore, a discussion of equilibrium of air should be separate with respect to dry air and saturated air.

Equilibrium of Dry Air

The method used for determining the equilibrium of air will be the parcel method, wherein a parcel of air is lifted and then compared to the surrounding air to determine its equilibrium. The dry adiabatic lapse rate is always used as a reference to determine the stability or instability of dry air.

ABSOLUTE INSTABILITY.–Consider a column of air in which the actual lapse rate is greater than the dry adiabatic lapse rate (the actual lapse rate is to the left of the dry adiabatic lapse rate on the Skew T diagram). (See fig. 3-8.) If the parcel of air at point A were displaced upward to point B, it would cool at the dry adiabatic lapse rate. Upon arriving at point B, it would be warmer than the surrounding air. The parcel would therefore have a

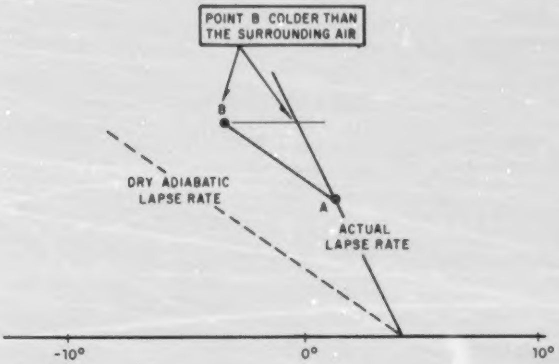

POINTS B (DRY ADIABATIC) AND B' (MOIST ADIABATIC) WARMER THAN SURROUNDING AIR
B
B'
ACTUAL LAPSE RATE
A
DRY ADIABATIC LAPSE RATE
SATURATION ADIABATIC LAPSE RATE
-10°
0°
10°

AG.399

Figure 3-8.–Absolute instability (any degree of saturation).

tendency to continue to rise, seeking air of its own density. Consequently the column would be unstable. From this, the rule is established that if the lapse rate of a column of air is greater than the dry adiabatic lapse rate, the column is in a state of ABSOLUTE INSTABILITY. The term "absolute" is used because this applies whether the air is dry or saturated, as is evidenced by displacing upward a saturated parcel of air from point A along a saturation adiabat to point B′. The parcel is more unstable than if displaced along a dry adiabat.

STABILITY.–Consider a column of dry air in which the actual lapse rate is less than the dry adiabatic lapse rate (the actual lapse rate is to the right of the dry adiabatic lapse rate on the Skew T diagram). (See fig. 3-9.) If the parcel at point A were displaced upward to point B, it would cool at the dry adiabatic lapse rate, and upon arriving at point B it would be colder than the surrounding air. It would therefore have a tendency to return to its original level. Consequently, the column of air would be stable.



AG.400

Figure 3-9.–Stability (dry air).

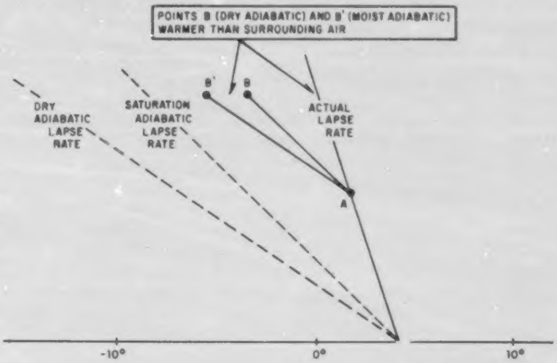
From this, the rule is established that if the actual lapse rate of a column of DRY AIR is less than the dry adiabatic lapse rate, the column is stable.

NEUTRAL STABILITY.–Consider a column of DRY AIR in which the actual lapse rate is equal to the dry adiabatic lapse rate. The parcel would cool at the dry adiabatic lapse rate if displaced upward. It would at all times be at the same temperature and density as the surrounding air, and would have a tendency neither to return to nor to move farther away from its original position. The column of dry air therefore, would be in a state of NEUTRAL STABILITY.

Equilibrium of Saturated Air

When saturated air is lifted, it cools at a rate different from that of dry air. This is due to release of the latent heat of condensation, which is absorbed by the air. The rate of cooling of saturated air is known as the saturation adiabatic lapse rate. This rate is used as a reference for determining the equilibrium of saturated air.

ABSOLUTE STABILITY.–Consider a column of air in which the actual lapse rate is less than the saturation adiabatic lapse rate (the actual lapse rate is to the right of the saturation adiabatic lapse rate on the Skew T diagram). (See fig. 3-10.) If the parcel of saturated air at point A were displaced upward to point B, it would cool at the saturation adiabatic lapse rate, and upon arriving at point B it would be colder than the surrounding air.


POINTS B (DRY ADIABATIC) AND B' (MOIST ADIABATIC) WARMER THAN SURROUNDING AIR
B'
B
DRY ADIABATIC LAPSE RATE
SATURATION ADIABATIC LAPSE RATE
ACTUAL LAPSE RATE
A
-10°
0°
10°

AG.401

Figure 3-10.–Absolute stability (any degree of saturation).

The layer therefore would be in a state of ABSOLUTE STABILITY. From this, the following rule is established: If the actual lapse rate for a column of air is less than the saturation adiabatic lapse rate, the column is absolutely stable. Dry air cools dry adiabatically and also would be colder than the surrounding air. Therefore, this rule applies to all air, as is evidenced when an unsaturated parcel of air is displaced upward dry adiabatically to point B′, where the parcel is more stable than the parcel displaced along a saturation adiabat.

INSTABILITY.–Consider now a column of air in which the actual lapse rate is greater than the saturation adiabatic lapse rate. (See fig. 3-11.) If a parcel of saturated air at point A were displaced upward to point B, it would cool at the saturation adiabatic lapse rate. Upon arriving at point B the parcel would be warmer than the surrounding air. For this reason, it would have a tendency to continue moving farther from its original position. The parcel therefore would be in a state of INSTABILITY. The following rule is applicable: If the actual lapse rate for a column of SATURATED AIR is greater than the saturation adiabatic lapse rate, the column is unstable.

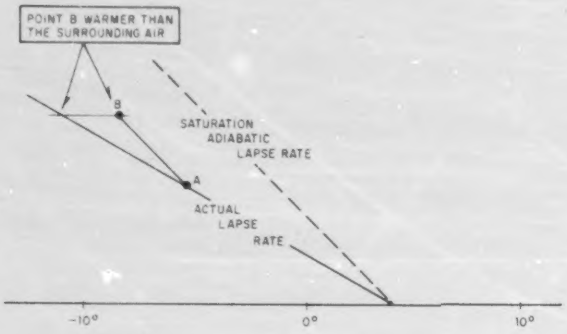

POINT B WARMER THAN THE SURROUNDING AIR
B
SATURATION ADIABATIC LAPSE RATE
A
ACTUAL LAPSE RATE
-10°
0°
10°
AG.402

Figure 3-11.—Instability (saturated air).

NEUTRAL STABILITY.–Consider a column of saturated air in which the actual lapse rate is equal to the saturation adiabatic lapse rate. A parcel of air displaced upward would cool at the saturation adiabatic lapse rate and would at all times be equal in temperature to the surrounding air. On that account, it would tend neither to move farther away from nor to return to its original level. It therefore would be in a state of NEUTRAL STABILITY. The rule for this situation is that if the actual lapse rate for a column of saturated air is equal to the saturation adiabatic lapse rate, the column is neutrally stable.

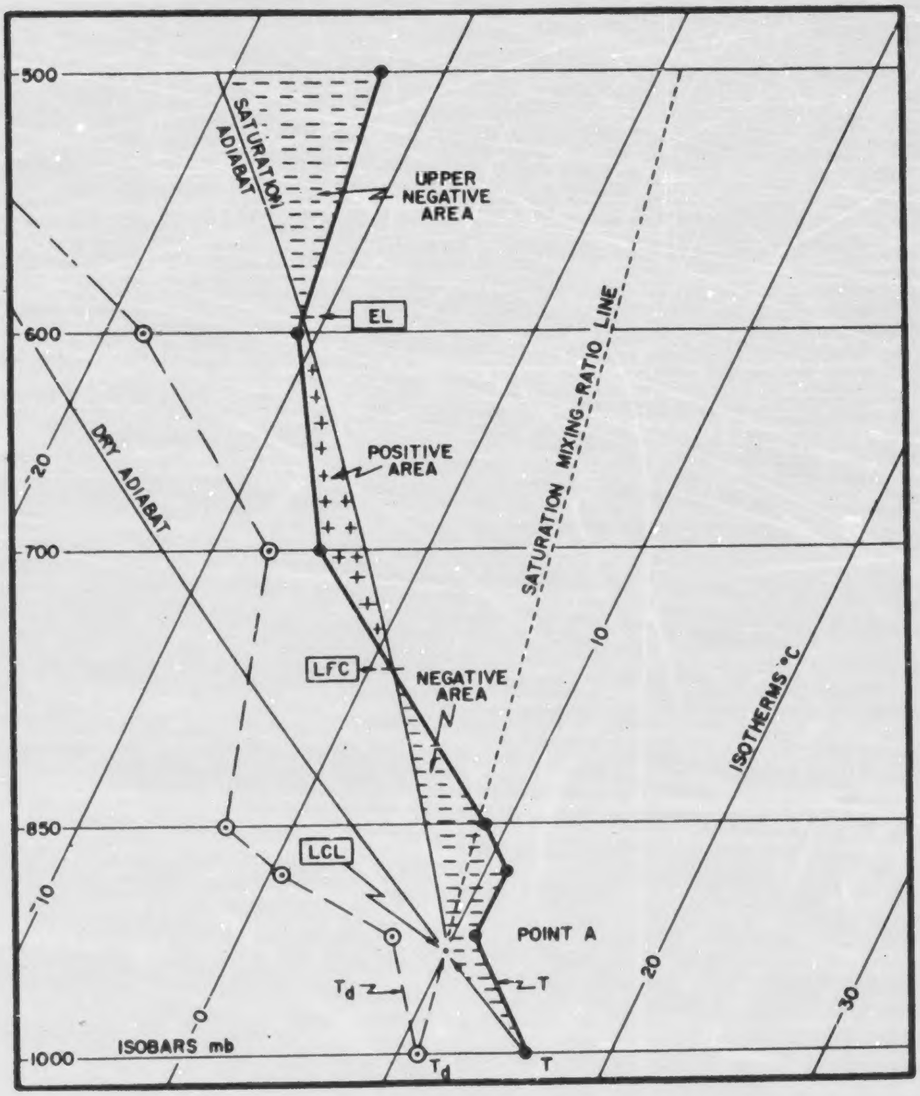
Conditional Instability

In the treatment of stability and instability so far, only air that was either dry or saturated was considered. Under normal atmospheric conditions natural air is unsaturated to begin with, but becomes saturated if lifted far enough. This presents no problem if the actual lapse rate for the column of air is greater than the dry adiabatic lapse rate (absolutely unstable) or if the actual lapse rate is less than the saturation adiabatic lapse rate (absolutely stable). However, if the lapse rate for a column of natural air lies between the dry adiabatic lapse rate and the saturation adiabatic lapse rate, the air may be stable or unstable, depending upon the distribution of moisture. When the actual lapse rate of a column of air lies between the saturation adiabatic lapse rate and the dry adiabatic lapse rate, the equilibrium is termed CONDITIONAL INSTABILITY, because the stability is conditioned by the moisture distribution. The equilibrium of this column of air is determined by the use of positive and negative energy areas. The determination of an area as positive or negative depends upon whether the environment is colder or warmer than the ascending parcel. Positive areas are conducive to instability; negative areas are conducive to stability.

TYPES OF CONDITIONAL INSTABILITY.–Conditional instability may be one of three types. The REAL LATENT type is a condition in which the positive area is larger than the negative area (potentially unstable). The PSEUDOLATENT type is a condition in which the positive area is smaller than the negative area (potentially STABLE). The stable type is a condition in which there is no positive area.

ENERGY AREAS FOR MECHANICAL LIFTING.–A negative area is the area on a Skew T diagram) bounded by the temperature curve; the dry adiabat from the surface point to the lifting condensation level (LCL); and the moist adiabat from the LCL to its intersection with the temperature curve (this point on the temperature curve is termed the level of free convection (LFC)). In figure 3-12, the negative area is shaded with dashed lines.

A positive area is the area (on a Skew T diagram) to the right of the temperature curve,

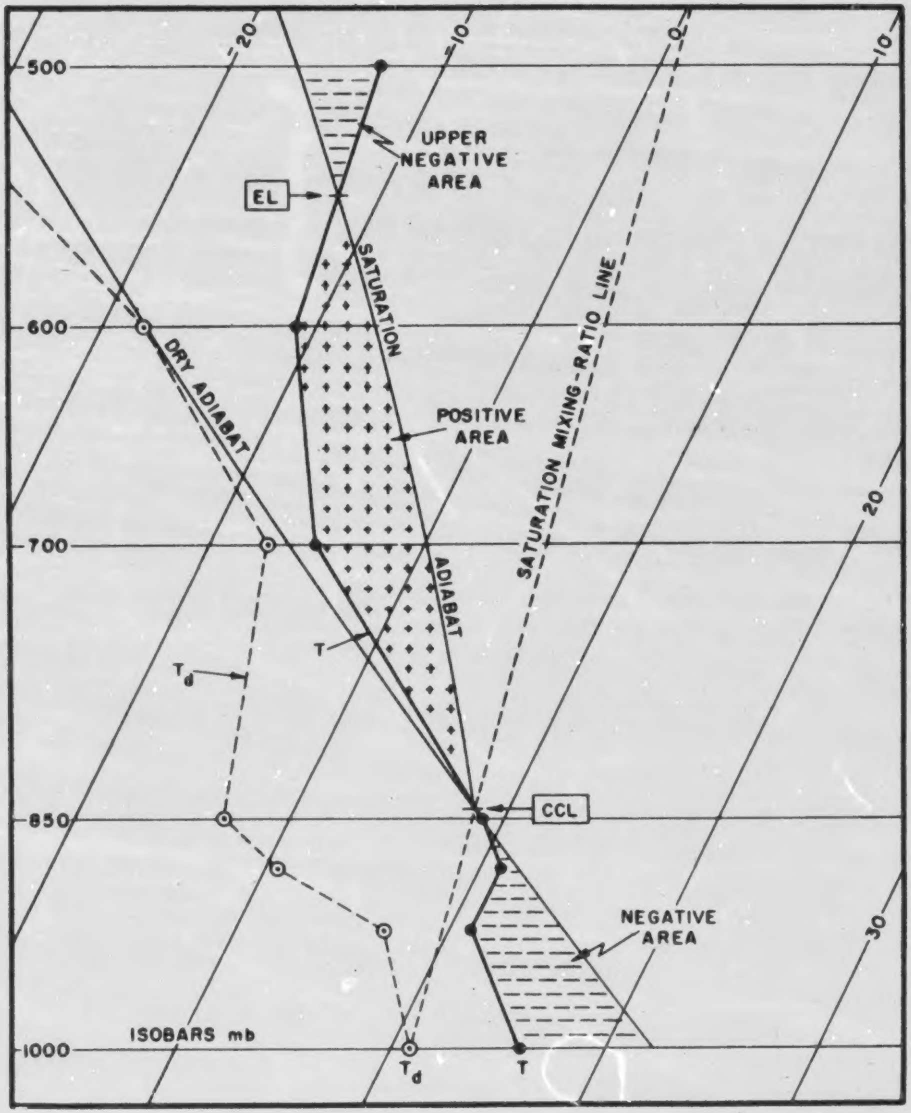

500
600
700
850
1000
ISOBARS mb
SATURATION ADIABAT
UPPER NEGATIVE AREA
EL
POSITIVE AREA
DRY ADIABAT
SATURATION MIXING-RATIO LINE
LFC
NEGATIVE AREA
LCL
POINT A
T_d
T
ISOTHERMS °C
-20
-10
0
10
20
30

AG.403

Figure 3-12.—LCL, LFC, and negative and positive areas—mechanical lifting.

bounded by the temperature curve and the saturation adiabat extended upward from the LFC. In figure 3-12, it is shaded with a cross pattern.

ENERGY AREAS FOR CONVECTIVE LIFTING.–If lifting by convection is expected, the negative area is determined by locating the intersection of the temperature curve, the average


-20
-10
0
10
500
600
700
850
1000
UPPER NEGATIVE AREA
EL
SATURATION ADIABAT
POSITIVE AREA
SATURATION MIXING-RATIO LINE
DRY ADIABAT
20
T
T_d
CCL
NEGATIVE AREA
30
ISOBARS mb
T_d
T

AG.404

Figure 3-13.–CCL, and negative and positive areas–convective lifting.

mixing ratio, and a dry adiabat. The average mixing ratio is chosen because more than just surface parcels are involved in convective activity. The standard practice is to average the mixing ratio for a 100-mb stratum above the surface, or to average the mixing ratio for the moist surface layer when it is less than 100 millibars in vertical extent. In addition, each locality should add a small factor, determined locally, to allow for an increase in moisture; this is especially true on coastlines or near large rivers and lakes. This procedure insures use of a realistic moisture content for the lower layer (which will be thoroughly mixed in the convective process). This level is known as the convective condensation level (CCL). The area downward from this intersection and bounded by the temperature curve and the dry adiabat is the negative area. It represents the energy which must be supplied in order that a parcel of air will rise from the surface to a level where it will continue to rise without a supply of energy from an outside source. The intersection of the dry adiabat drawn from the CCL with the surface level determines the surface temperature necessary for free convection. Notice that in this situation the negative area is to the right of the temperature curve, whereas with mechanical lifting the negative area is to the left of the temperature curve. (See fig. 3-13.)

The positive area in a situation of convective lifting is the area to the right of the temperature curve, bounded by the temperature curve and the saturation adiabat extended from the intersection of the mixing ratio and the temperature curve; that is, the CCL. (See fig. 3-13.)

Autoconvection

AUTOCONVECTION is a condition which is started spontaneously by a layer of air when the lapse rate of temperature is such that density increases with elevation. For density to increase with altitude, the lapse rate must be equal to or exceed 3.42°C per 100 meters. (This is the AUTOCONVECTIVE LAPSE RATE.) An example of this condition is found to exist near the surface of the earth in a road mirage or a dust devil. These occur over surfaces which are easily heated, such as the desert, open fields, etc., and are usually found during periods of intense surface heating.

Convection Stability and Instability

In the discussion so far of convection stability and instability, PARCELS of air have been considered. Let us now examine LAYERS of air. A layer of air which is originally stable may become quite unstable due to moisture distribution if the entire layer is lifted.

Convective stability is the condition that occurs when the equilibrium of a layer of air, because of the temperature and humidity distribution, is such that when the entire layer is lifted, its stability is increased.

Convective instability is the condition of equilibrium of a layer of air occurring when the temperature and humidity distribution is such that when the entire layer of air is lifted, its instability is increased.

CONVECTIVE STABILITY.–Consider a layer of air whose humidity distribution is dry at the bottom and moist at the top. If the layer of air is lifted, the top and the bottom will cool at the same rate until the top reaches saturation. Thereafter, the top will cool less rapidly than the bottom. The top will cool saturation adiabatically; the bottom will still continue to cool dry adiabatically. The lapse rate for the layer then will decrease. The stability will increase.

The layer must be unstable at the beginning and may become stable when lifting takes place.

In a layer that is convectively stable, the equivalent potential temperature increases with elevation.

CONVECTIVE INSTABILITY.–Consider a layer of air in which the air at the bottom is moist and the air at the top of the layer is dry. If this layer of air is lifted, the bottom and the top will cool dry adiabatically until the lower portion is saturated. The lower part will then cool saturation adiabatically while the top of the layer is still cooling dry adiabatically. The lapse rate then begins to increase, the instability increasing.

In a layer of convectively unstable air, the equivalent potential temperature DECREASES with elevation.

In order to determine the convective stability or instability of a layer of air, you should first know why you expect the lifting of a whole layer. The obvious answer is an orographic barrier or a frontal surface. Next, determine how much lifting is to be expected and at what level does it commence, for you need not necessarily have to lift a layer of air close to the surface of the earth. The amount of lifting will, of course, depend on the situation at hand. In determining the convective stability or instability of a layer of air at a particular locality proceed as follows:

1. Lift the lowest end of the lapse rate along the appropriate adiabat(s) (dry, moist, or dry then moist upon saturation) for a predetermined number of millibars.
2. Lift the upper end of the lapse rate along the appropriate adiabat(s) for an equal number of millibars.
3. Connect the upper to the lower point thus formed with a straight line, representing the new lapse rate.

Stability Determination from Existing Lapse Rate

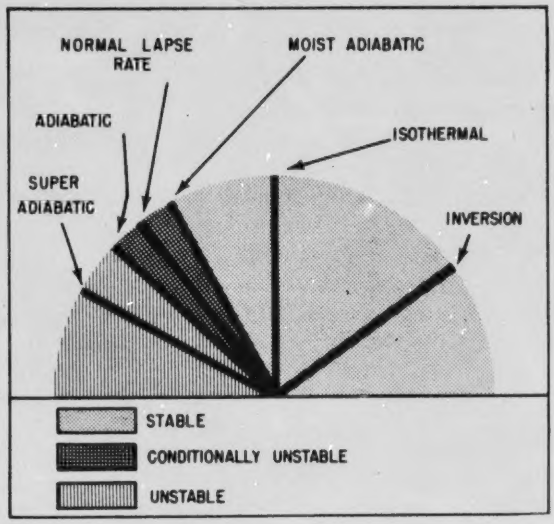
We can very simply test the stability conditions of the plotted sounding by observing the observed temperatures and lapse rates in reference to superimposed lines representing the dry and moist adiabatic lapse rates on the Skew T diagram.

For instance, if we observe that the lapse rate on the actual sounding is to the right of the moist adiabatic rate, the air is absolutely stable. If it lies between the moist and dry adiabatic lapse rates, its stability is dependent on the moisture present, and it is called conditionally stable. If the lapse rate is greater than the dry adiabatic lapse rate, we have absolute instability in the air and can expect vertical currents to cause turbulence in that area.

Figure 3-14 illustrates the varying degrees of air stability which are directly related to the rate at which the temperature changes with height.

Determining Bases of Convective Type Clouds

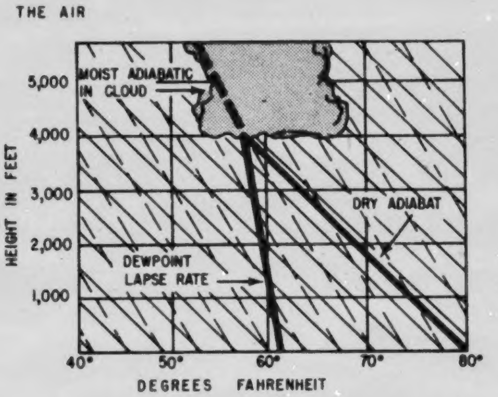
We have seen from our foregoing discussion in an earlier section of this chapter that moisture is important in determining certain stability conditions in the atmosphere. We know, too, that the difference between the temperature and the dewpoint is an indication of the relative humidity, and that when the dewpoint and the temperature are the same, the air is saturated and some form of condensation cloud may be expected. This lends itself to a means of estimating the height of the base of clouds formed by surface heating; that is, cumuliform type clouds, when the surface temperature and dewpoint are known. We know that the dewpoint will decrease in temperature at the rate of 1°F per 1,000 feet during a lifting process. The ascending parcel in the convective current will experience a decrease in temperature of about 5 1/2°F per 1,000 feet. Thus the dewpoint and the temperature approach each other at the rate of 4 1/2°F per 1,000 feet. As an example, consider the surface temperature to be 80°F and the surface dewpoint 62°F, a difference of 18°F. This difference, divided by the approximate rate that the temperature approaches the dewpoint (4 1/2°F per 1,000 ft) indicates the approximate height of the base of the clouds caused by this lifting process (18 ÷ 4 1/2 =



AG.405

Figure 3-14.—Degrees of stability in relation to temperature changes with height.

4,000 feet). This is graphically shown in figure 3-15.



AG.406

Figure 3-15.—Determination of cloud's base when the dewpoint and temperature are known.

The Aerographer's Mate should remember that this method cannot be applied to all cloud types, but is limited to clouds formed by convection currents, such as summertime cumulus clouds, and only in the locality where the clouds form. It is not valid around maritime or mountainous areas.

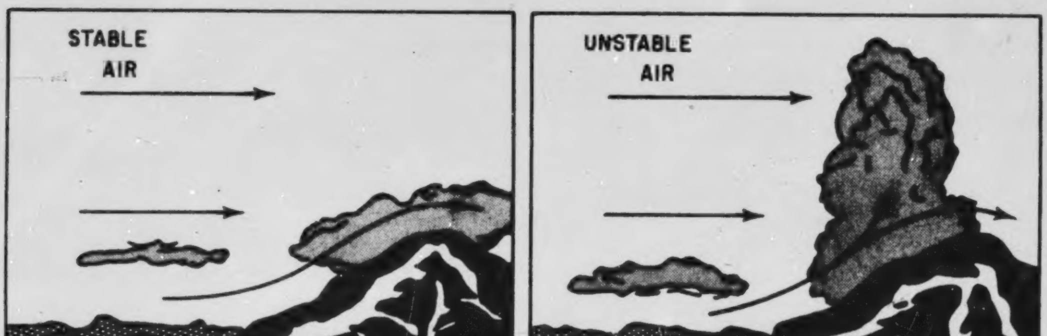
Stability in Relation to Cloud Type

When a cloud is formed, the stability of the atmosphere helps to determine the type of cloud formed. For example, if the air is very stable and it is being forced to ascend the side of a mountain (fig. 3-16), the cloud formed will be layerlike with little vertical development and with little or no turbulence. If, however, the air is unstable in the situation, the clouds formed would have vertical development and turbulence would be expected with them. The base of this type of cloud would be determined by mechanical lifting and by the LCL.

ELECTROMAGNETIC RADIATION

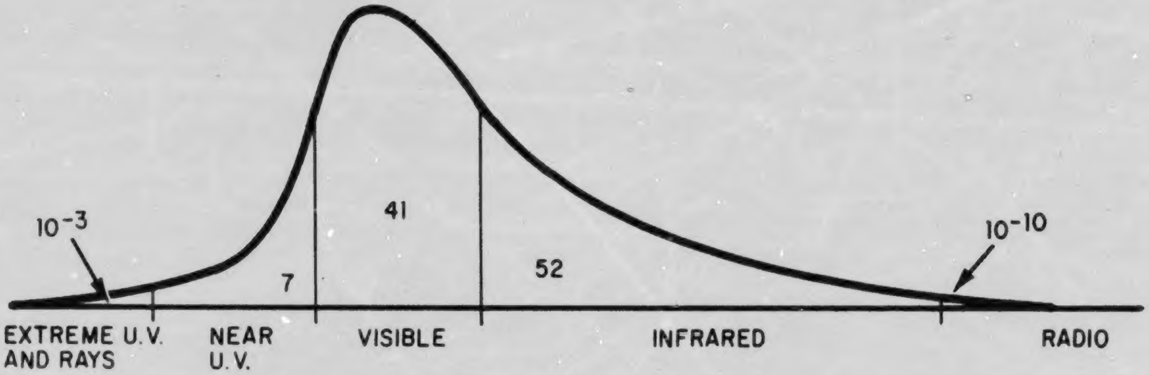
Electromagnetic energy as related to light and heat were discussed in AG 3 & 2, NavTra 10363-D. The various types of electromagnetic energy comprising the electromagnetic spectrum are radio, infrared, visible, ultraviolet, x-rays, gamma rays, and cosmic rays. (See fig. 3-17.)

The term spectrum, as it is used here, refers to the whole range of electromagnetic radiations. Although the basic nature of electromagnetic waves is the same (for example, they all travel at 186,000 miles per second), they do differ in their wavelength. The wavelength determines

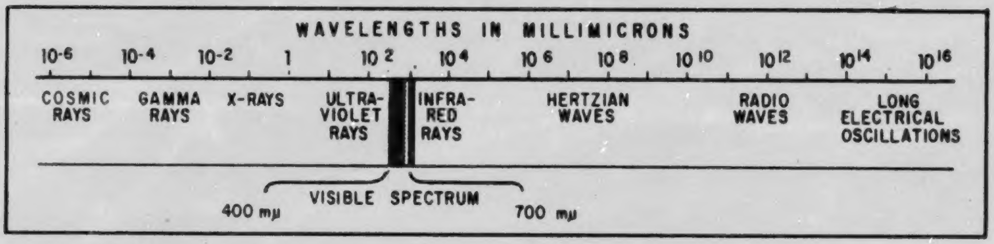


AG.407

Figure 3-16.—Illustration showing that very stable air retains its stability even when it is forced upward, forming a flat cloud. Air which is potentially unstable when forced upward becomes turbulent and forms a towering cloud.



SCHEMATIC DIAGRAM OF THE DISTRIBUTION OF ENERGY IN THE SOLAR SPECTRUM. (NOT TO SCALE). THE NUMBERS ARE PERCENTAGES OF THE SOLAR CONSTANT. THE FIGURE FOR THE RADIO ENERGY IS FOR THE OBSERVED BAND FROM 15 TO 30,000 MHZ.


WAVELENGTHS IN MILLIMICRONS
10^{-6}
10^{-4}
10^{-2}
1
10^{2}
10^{4}
10^{6}
10^{8}
10^{10}
10^{12}
10^{14}
10^{16}
COSMIC RAYS
GAMMA RAYS
X-RAYS
ULTRA-VIOLET RAYS
INFRA-RED RAYS
HERTZIAN WAVES
RADIO WAVES
LONG ELECTRICAL OSCILLATIONS
VISIBLE SPECTRUM
400 mμ
700 mμ

AG.408

Figure 3-17.—Electromagnetic spectrum.

whether they will be categorized as infrared, ultraviolet, radio, etc.

During periods of significant solar activity, radiative energy increases of as much as 1,000 percent may occur in various frequency ranges (radio, ultra-violet, and x-ray). This energy contribution to the total emission over the entire spectrum is very, very small because these frequency ranges account for such a small part of the total energy, as shown in the upper part of figure 3-17. However, this energy increase in these particular frequencies is extremely significant in its effect on some of man's activities; in particular, it is significant to military operations which are concerned with satellite surveillance and with communications. Consequently, some military units are currently monitoring solar activity, and reporting, analyzing, and predicting solar activity. The U.S. Navy has developed a system for space environmental monitoring and related forecasting technology—SOLRAD-HI (Solar Radiation-High Altitude) Satellite Experiment. Launch of the satellites is scheduled for the mid-70's; a Naval Weather Service Solar Forecast Center is scheduled for establishment in the latter half of the 70's. From this, it should be apparent that the Aerographer's Mate must have some knowledge of solar activity and its effects. This chapter will provide some general information on the sun, its atmosphere, and some of the more notable features of the solar disk. A later chapter, Special Observations and Forecasts, will provide additional information.

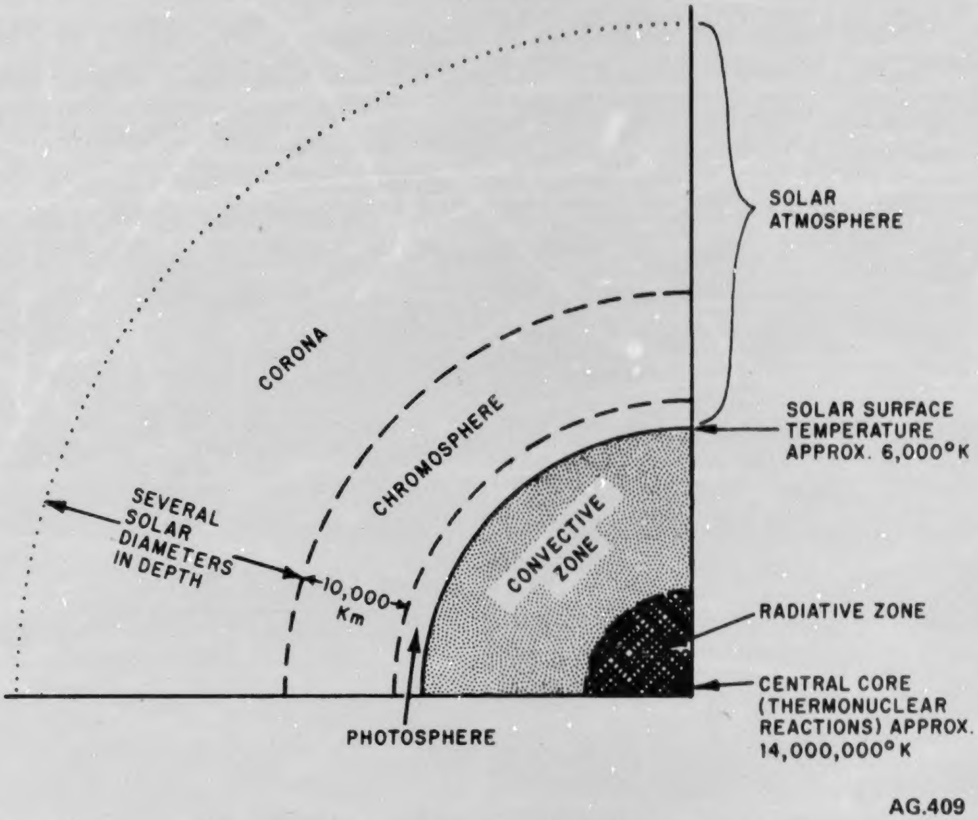

SOLAR ATMOSPHERE
CORONA
CHROMOSPHERE
SOLAR SURFACE TEMPERATURE APPROX. 6,000°K
SEVERAL SOLAR DIAMETERS IN DEPTH
10,000 Km
CONVECTIVE ZONE
RADIATIVE ZONE
CENTRAL CORE (THERMONUCLEAR REACTIONS) APPROX. 14,000,000° K
PHOTOSPHERE
AG.409

Figure 3-18.—One-quarter cross-section depicting solar structure.

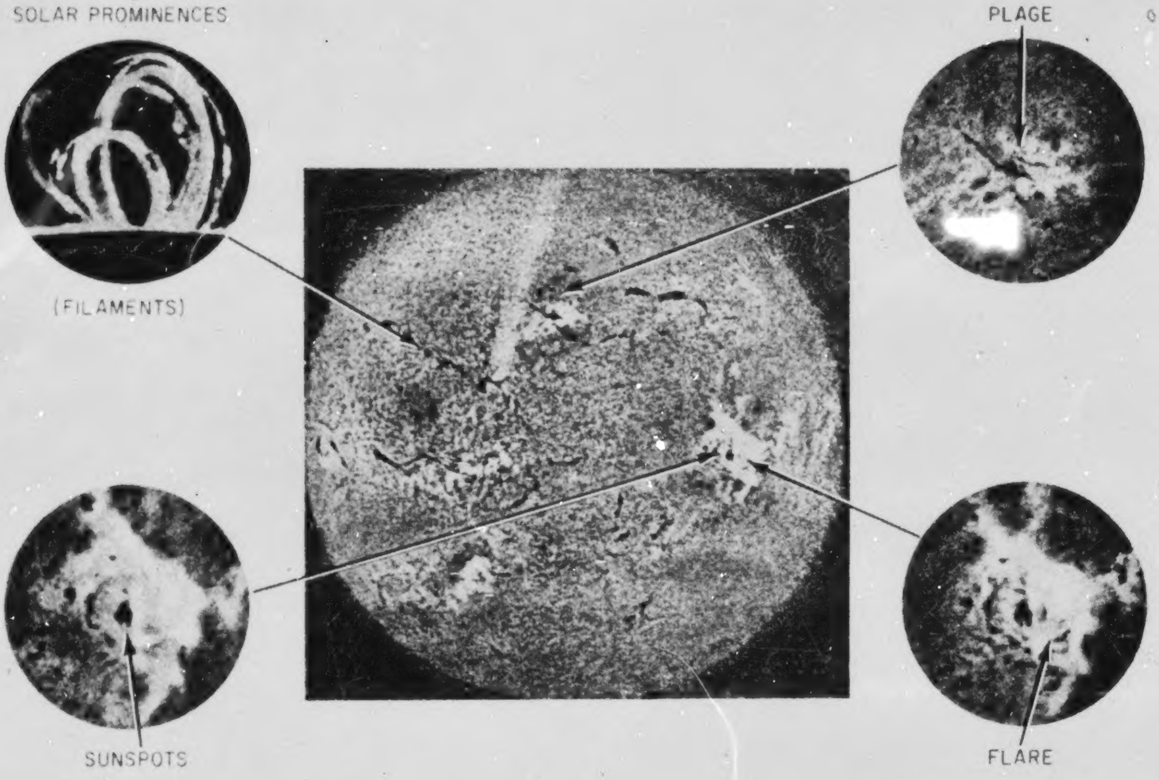
FEATURES OF THE SOLAR DISK

The term solar disk refers to a visual image of the outer surface of the sun as observed from outside regions.

The sun may be described as a globe of gas heated to incandescence by thermonuclear reactions from within the central core. (See fig. 3-18.)

The tremendous heat (or energy) generated from within the sun's core is transported by the radiative transfer of photons (a measurement of gamma radiation), which bounce from atom to atom similar to bouncing balls through the radiative zone. Within the convective zone, which extends very nearly to the sun's surface, the heated gases are raised buoyantly upwards with some cooling occurring and subsequent convective action as within the earth's atmosphere. The gases are cooled to approximately 6,000 K (Kelvin or absolute) at the sun's surface.

The main body of the sun, although composed of gases, is opaque and has a well-defined visible surface referred to as the photosphere. This is the source we see from which all the light and heat of the sun is radiated. Above the photosphere is a more transparent gaseous layer referred to as the chromosphere with a thickness of about 10,000 km. It is hotter than the photosphere. Above the chromosphere is the corona, a low density high temperature region, which is extended far out into interplanetary space by the solar wind—a steady outward streaming of the coronal material. Much of the



AG.410

Figure 3-19.—Features of the solar disk.

solar x-ray and solar radio emission originates in the corona.

Within the solar atmosphere certain more transient phenomena (referred to as solar activity) occur just as cyclones, frontal systems, and thunderstorms occur within the atmosphere of the earth. This activity may consist of the phenomena discussed in the following paragraphs which collectively describe the features of the solar disk. (See fig. 3-19.)

Solar Prominences/Filaments

Solar prominences/filaments are injections of gases from the chromosphere into the corona. They appear as great clouds of gas, sometimes resting on the sun's surface and at other times floating free with no visible connection. When viewed against the solar disk as illustrated in figure 3-19, they appear as long dark ribbons and are called filaments; when viewed against the solar limb, they appear bright and are called prominences. They display a variety of shapes, sizes, and activity which defy general description; they have a fibrous structure and appear to resist solar gravity. They may extend 30,000 to 40,000 km above the chromosphere. The more active types appear hotter than the surrounding atmosphere with temperatures near 10,000,000° K.

Sunspots

Sunspots are regions of strong localized magnetic fields and indicate relatively cool areas in the photosphere. They appear darker than their surroundings and may appear singly or in more complicated groups dominated by larger spots near the center. (See fig. 3-19.)

Sunspots begin as small dark areas known as pores. These pores develop into full-fledged spots in a few days, with maximum development occurring in about 1 to 2 weeks. Decaying of the sunspots consists of the spot shrinking in size with an accompanying decrease in the magnetic field. This life cycle may consist of a few days for small spots to near 100 days for larger groups. The larger spots normally measure about 120,000 km. Sunspots appear to have cyclic variations in intensity, varying through a period of about 8 to 17 years. Variation in number and size will occur throughout the sunspot cycle. As a cycle commences, a few spots are observed at high latitudes of both solar hemispheres, increasing in size and number. They gradually drift equatorward as the cycle progresses, reaching a maximum in about 4 years. After this period, decay will set in and near the end of the cycle only a few spots appearing in the lower latitudes (5° to 10°) will be left.

Plages

Plages such as those indicated in figure 3-19 are large irregular bright patches which surround sunspot groups. They normally appear in conjunction with solar prominences or filaments and may be systematically arranged in radial or spiral patterns. Plages are features of the lower chromosphere and often completely or partially obscure an underlying sunspot.

Flares

Solar flares are perhaps the most spectacular of the eruptive features associated with solar activity. (See fig. 3-19.) They appear as flecks of light which suddenly appear near activity centers, appearing instantaneously as though a switch were thrown. They rise sharply to peak brightness in a few minutes, then decline more gradually. The number of flares may increase rapidly over an area of activity. Small flarelike brightenings are always in progress during the more active phase of activity centers. In some instances flares may take the form of prominences, violently ejecting material into the solar atmosphere and breaking into smaller highspeed blobs or clots. Flare activity appears to vary widely between solar activity centers. The greatest flare productivity seems to be during the week or 10 days when sunspot activity is at its maximum.

Flares are classified according to size and brightness. In general, the higher the importance classification, the stronger the geophysical effects. Some phenomena associated with solar flares have immediate effects; others, delayed effects (15 minutes to 72 hours after flare). Such phenomena as communications problems and plans for moving toward a solution to these problems are briefly discussed in chapter 16 of this manual.

CHAPTER 4

ATMOSPHERIC CIRCULATION

The direction of motion air follows within the atmosphere may be either horizontal or vertical, or it might move both horizontally and vertically. It is very important for the Aerographer's Mate to be able to determine the types of motion which exist and understand the effects this motion may have on weather changes within the atmosphere. It must be remembered that the existing atmospheric circulations (air movements) will continually cause changes in pressure, temperature, and moisture content as well as various other factors comprising the total atmosphere.

BASIC WIND THEORY

The basic rules pertaining to the relationships between pressure gradient and wind were discussed in chapter 5 of AG 3 & 2, NavTra 10363-D, along with a brief discussion of geostrophic and gradient wind.

In this section we will take a closer look at these circulations, the forces involved, and the effect they have on the general circulation.

GEOSTROPHIC WIND

GEOSTROPHIC WIND is a steady horizontal air motion along straight, parallel isobars in an unchanging pressure field, with gravity the only external force, and in a direction perpendicular to that in which the Coriolis force and the pressure gradient force as acting equally and oppositely.

In the above definition you may have noted that gravity is the only external force. Thus, it should be apparent that in order to have geostrophic wind, there must be no frictional force involved. Let us consider, then, a parcel of air from the time it begins to move until it develops into a geostrophic wind.

As soon as a parcel of air starts to move due to the pressure gradient force, the Coriolis force begins to deflect it from the direction of the pressure gradient force. The Coriolis force is the apparent force exerted upon a parcel of air due to the rotation of the earth. This force acts to the right of the path of motion of the parcel of air in the Northern Hemisphere and to the left in the Southern Hemisphere. It always acts at right angles to the direction of motion. In the absence of friction, the Coriolis force will change the direction of motion of the parcel until the Coriolis force and the pressure gardient force are in balance. When the two forces are equal and opposite, the wind will blow parallel to the straight isobars. It may be well to remember that the Coriolis force affects the direction but not the speed of the motion of the air. Normally, Coriolis force would not be greater than the pressure gradient force. In the case of supergradient winds, Coriolis force may be greater than the pressure gradient force. This will cause the wind to be deflected more to the right in the Northern Hemisphere, or toward higher pressure.

Figure 4-1 shows from left to right the changes, in three stages, that take place from the time the parcel of air begins to move until the final development of a geostrophic wind. The vectors in the illustration indicate direction only, not magnitude.

Geostrophic wind is dependent to a certain extent upon the density of the atmosphere and the latitude. If the density and the pressure gradient remain constant and the latitude

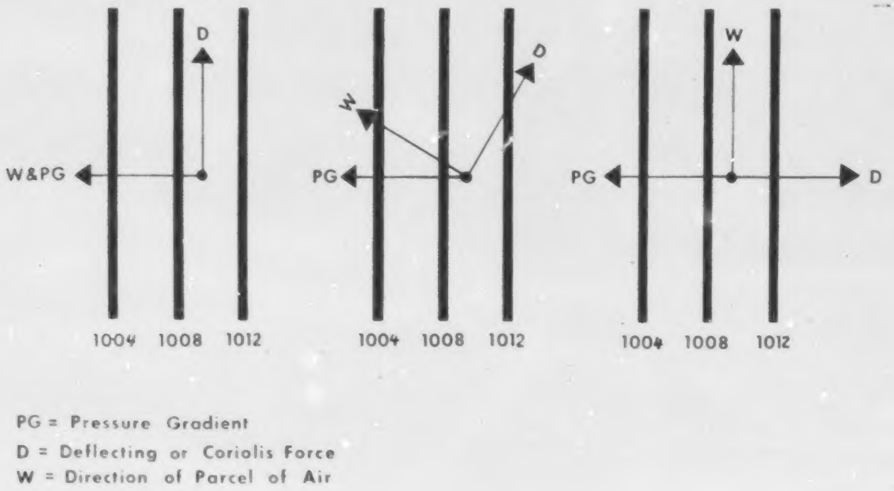


Figure 4-1.—Geostrophic wind.

increases, the wind speed decreases. On the other hand, if the latitude decreases, the wind speed increases. If the density and the latitude remain constant and the pressure gradient decreases, the wind speed decreases. If the pressure gradient and the latitude remain constant and the density decreases, the wind speed increases. If the density increases, the wind speed decreases.

True geostrophic wind is seldom observed in nature, but the conditions are closely approximated on upper-level charts.

GRADIENT WIND

GRADIENT WIND is the wind that flows parallel to the curved isobars in an UNCHANGING PRESSURE FIELD, when the centrifugal, Coriolis, and pressure gradient forces balance. As in the case of geostrophic wind, there is no frictional force acting. True gradient winds are rarely observed in nature, but rather a mutual coexistence of supergradient and subgradient winds. (In the case of supergradient conditions, the wind would be too strong for the existing pressure gradient and would be evidenced by a component of the wind across the isobars toward higher pressure.)

Centrifugal Force

When a parcel of air moves in a curved path, a force acts on the parcel of air and tends to throw the parcel outward from the center about which it is moving. This is CENTRIFUGAL FORCE.

Movement of Air Parcels Around Anticyclones

The movement of gradient winds around anticyclones is affected in a certain manner by the pressure gradient force, the centrifugal force, and the Coriolis force. As you would expect, the pressure gradient force acts from high to low pressure; and the Coriolis force acts opposite to the pressure gradient force and at right angles to the direction of movement of the parcel. The centrifugal force acts at right angles to the path of motion and outward from the center about which the parcel is moving. (See fig. 4-2.) In this case, the pressure gradient force and the centrifugal force balance the Coriolis force. (It may be expressed in the following manner: PG + CF = D.)

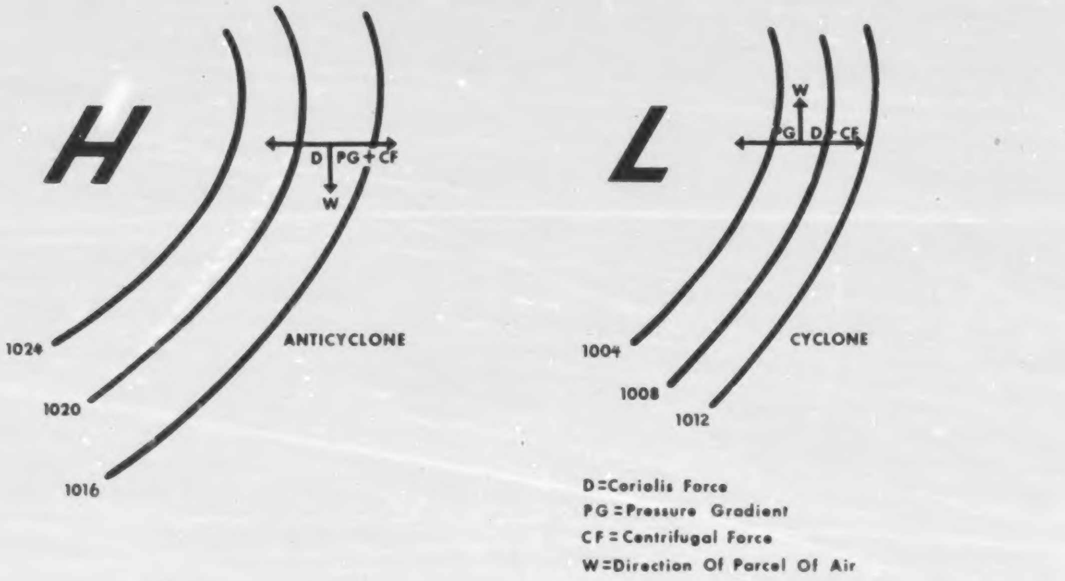


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Figure 4-2.—Forces acting on pressure systems.

Movement of Air Parcels Around Cyclones

As in the case of anticyclones, gradient winds around cyclones are affected by the pressure gradient force, the centrifugal force, and the Coriolis force, but the balance of the forces is different. (See fig. 4-2.) In a cyclonic situation the pressure gradient force is balanced by the Coriolis force and the centrifugal force. (It may be expressed in the following manner: PG = D + CF.)

Centrifugal force acts WITH the pressure gradient force when the circulation is anticyclonic and AGAINST the pressure gradient force when the circulation is cyclonic. Therefore, wind velocity will be greater with an anticyclone than with a cyclone of the same isobaric spacing.

Conservation of Angular Momentum

The apparent deflection of a parcel of air moving from south to north or from north to south may be explained on the basis of the law of CONSERVATION OF ANGULAR MOMENTUM. This law states that the angular velocity of a parcel multiplied by the radius of rotation squared, is equal to a constant. In equation form this may be written $\omega r^2 = R$, where ω is the angular velocity of the parcel. (The angular velocity of a parcel is the value of the angle through which a parcel moves per unit time; it is not linear distance.) The radius of rotation, that is, the perpendicular distance the parcel is from the axis of rotation of the earth, is represented by r. The constant is represented by R.

If a parcel of air moves northward in the Northern Hemisphere, its radius of rotation decreases. If r becomes smaller, ω must become larger in order that the product of r^2 and ω remain the same. If ω becomes larger, the parcel will have a tendency to move eastward relative to the earth's surface. That this actually happens can be demonstrated on a globe. For example, let us assume that a parcel of air is moving northward along a particular longitude at point A in figure 4-3. By the time the longitude has reached point A′ the parcel of air has moved to B because the angular velocity must increase due to the decrease of the radius of rotation. The distance from A to A′ must equal the distance from B to C. C is the point the parcel would have reached if the earth were stationary.

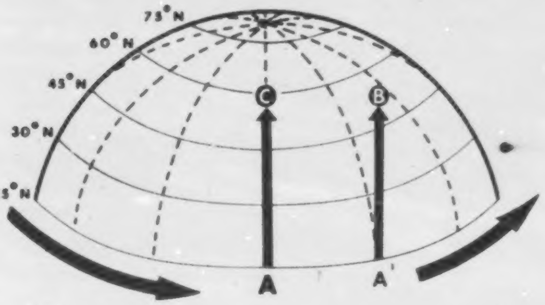

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Figure 4-3.—Parcel of air moving northward in the Northern Hemisphere.

If a parcel of air moves southward in the Northern Hemisphere, the radius of rotation increases and the angular velocity decreases. However, the parcel of air is apparently still deflected to the right.

In the Southern Hemisphere the apparent deflection of a northward or southward moving parcel of air is to the left.

Variations

In view of the foregoing discussion of the movement of air parcels around cyclones and anticyclones, it can be seen that with the same density, pressure gradient, and latitude, the wind will be weaker around a low-pressure cell than a high-pressure cell. This can also be seen from a comparison of the two types of gradient winds with the geostrophic wind. The wind we observe on a synoptic chart is usually stronger around low cells than high cells because the pressure gradient is usually stronger around the low-pressure cell.

In summary, under given conditions of latitude, density, and pressure gradient, the geostrophic wind is stronger than the gradient wind around a low and is weaker than a gradient wind around a high. It is for this reason that isobar spacing and contour spacing, when the flow is curved, differ from that determined by a geostrophic wind scale. It should be borne in mind that if the flow under consideration is around a high-pressure cell, the isobars will be farther apart than indicated by the geostrophic wind scale. And, if the flow under consideration is around a low-pressure cell, the isobars will be closer together than indicated by the geostrophic wind scale.

Geostrophic and Gradient Wind Scales

There are a number of scales available for measuring both geostrophic and gradient flow of both surface and upper air charts. For a detailed discussion of the theory and construction of these scales consult Meteorological Wind Scales, NavWeps 50-1P-551. An explanation of the use of these scales is covered in chapter 6 of this training manual.

Some of the weather plotting charts in use in the Naval Weather Service have geostrophic wind scales printed on them for both isobaric and contour spacing. A number of other scales are available from other publications and sources.

The two common scales in general usage are for sea level (4-mb isobars) and the pressure contour scale. The first is a scale used in determining the geostrophic wind speed, or isobar spacing on surface (sea level) charts, and the other is used for determining the geostrophic wind speed for contour spacing on constant/pressure charts. In tropical regions, the geostrophic wind scales become less reliable due to the fact that pressure gradients are generally rather weak.

CYCLOSTROPHIC WIND

In some atmospheric conditions, the radius of rotation becomes so small that the centrifugal force becomes quite great in comparison with the Coriolis force. This is particularly true in low latitudes where the Coriolis force is quite small to begin with. In this case, the pressure gradient force is nearly balanced by the centrifugal term alone. When this occurs, the wind is said to be cyclostrophic in flow. By true definition, a cyclostrophic wind exists when the pressure gradient force is balanced by the centrifugal force alone.

This exact situation rarely exists, but is so nearly reached in some situations that the small Coriolis effect is neglected and the flow is said to be cyclostrophic. Winds in a hurricane or typhoon, and the winds around a tornado are thought to be cyclostrophic.

FRICTIONAL EFFECTS

The surface wind is usually about two-thirds of the geostrophic wind. This reduction in wind speed is caused by friction. Since the Coriolis force varies with the speed of the wind, a reduction in the wind speed by friction means a reduction in the Coriolis force. This results in a momentary disruption of the balance and when the new balance, including friction is reached, the air blows at an angle across the isobars from high pressure to low pressure. The angle varies from 10° over the ocean to as much as 45° over rugged terrain. Frictional effects on the air are greatest near the ground, but the effects are also carried aloft by turbulence. Surface friction is effective in slowing down the wind up to about 1,500 to 3,000 feet above the ground. Above this level the effect of friction decreases rapidly and may be considered negligible for all parctical purposes. Therefore, air at about 3,000 feet or more above the ground usually tends to flow parallel to the isobars.

THE GENERAL CIRCULATION

The interplay of the factors which accomplish the balance of heat in the atmosphere is random, though always in response to cause. The atmosphere as a whole is constantly in the process of trying to establish equilibrium. In this equilibrium-seeking process, a general pattern of windflow is established—the general circulation.

The general circulation of the atmosphere is precisely that. Irregularities within that circulation are the rule rather than the exception and for this reason the general circulation is not covered at great length.

THEORY

The basis for the primary circulation is thermally driven motion of air expressed in terms of a solenoidal field. The earth's atmosphere is a huge, inefficient heat engine converting potential energy as represented by heat differences into kinetic energy of motion. The wide equatorial belt between latitudes 35° N and 35° S, representing over one-half the earth's surface, with a net gain of heat, and the higher latitudes with a net loss of heat, superimposed by a mean atmosphere where temperature decreases with elevation; adiabatic contribution of temperature; the expansion of the atmosphere over equatorial regions; and the contraction of the atmosphere over the poles form a perfect situation for the formation of a solenoidal field.

A solenoid is a tube formed in the atmosphere by the intersection of isotimic surfaces (isotimic surfaces are surfaces in space on which the value of a given property is everywhere equal) of two scalar quantities. In figure 4-4, the solenoidal field is representative of the mean conditions in the Northern Hemisphere. (Note the concentration of solenoids in the middle latitudes.)

A solenoid in figure 4-4 is formed by the intersection of isolines of specific volume and isobars and is shown by the enclosed area a, b, c, and d. It can be mathematically shown that the work done by a unit mass of air in a circulation arising from solenoid-producing thermal processes can be determined from the solenoidal field. The acceleration of such a circulation depends on the number of solenoids in the atmosphere. Without solenoids, the circulation comes to a standstill. The concept of a circulation arising from the formation of solenoids is not limited to the large-scale general circulation, but it has equal application in explaining the formation of secondary and tertiary circulations as well.

Solenoids have little application in everyday meteorological work; however, they have great application in research work, when new concepts are formulated or new or improved forecasting methods are devised. Solenoidal representation has particular application in numerical forecasting work when dealing with the fields of pressure, density, and temperature in the atmosphere. In connection with this particular application, two other terms may prove of value to the Aerographer's Mate because of the ever-increasing use of them. They are: baroclinity and barotropy. A BAROCLINIC field is one where the pressure/density and temperature field are nonparallel; that is, it is a solenoidal

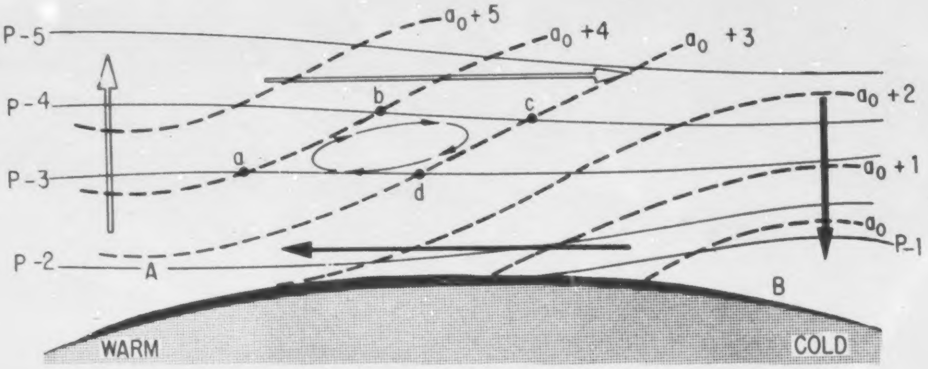

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Figure 4-4.—A solenoidal field.

field with pressure/density and temperature as coordinates. A BAROTROPIC field is one where isovalues of pressure/density and temperature are parallel to each other; that is, it contains no solenoids.

The solenoidal presentation does not EXPLAIN the general circulation; it is only the BASIS for the general circulation. For an explanation of the general circulation with the zonal winds, the consequent formation of the centers of action, and the influence of migratory disturbances, the physics of motion must be considered. Advanced treatment of the physics of motion is given in chapter 3 of this training manual. The general circulation, secondary circulation with the centers of action, monsoon circulations, migratory disturbances, and tertiary circulations are adequately treated in chapter 5 of AG 3 & 2. The general circulation also includes the special circulations, consisting primarily of the jetstream and zonal flow. Zonal flow is the basis for the zonal index. Both special circulations, the jetstream and zonal flow, receive treatment later in this chapter.

SECONDARY CIRCULATIONS

Since the earth does not have a uniform surface, the general circulation, as represented by more or less continuous pressure belts around the earth at the various latitudes, is not a true picture of actual conditions. Certain effects cause an organized system of highs and lows known as centers of action. There are also closed circulations that migrate and play an important role in the daily and seasonal changes of the weather.

There are two factors which cause the pressure belts of the primary circulation to break up into closed circulations of the secondary circulations. From our previous discussion in chapter 2 of this training manual, we know that the earth is composed of both land and water surfaces. The surface temperature of the ocean changes very little during the year, while in winter the land is cold and in summer it is warm. In the winter, highs form over cold continents and the lows over adjacent oceans. The reverse is true in the summer except in the area over the warm water belt near Greenland which tends to maintain low pressure. This effect of the difference in heating and cooling of land and water surfaces is known as the thermal effect.

Circulation systems are also created by the interaction of wind belts or systems, or the variation in wind in combination with certain distributions of temperature and/or moisture. This is known as the dynamic effect. This effect rarely, if ever, operates alone, in creating secondary systems, as most of the systems are both created and maintained by a combination of the thermal and dynamic effects.

CENTERS OF ACTION

When diagrams showing the distribution of mean surface temperature are compared with those showing mean sea level pressure, it can readily be seen that the pressure belts of the primary circulation are broken up and take a more cellular structure. The break corresponds with regions showing differences in temperature from land to water surfaces. These cells which tend to persist in a particular area are called centers of action: that is, they are found at nearly the same location with somewhat similar intensity during the same month each year.

There is a permanent belt of relatively low pressure along the Equator and another deeper ring of low pressure around the Antarctic. Permanent belts of high pressure largely encircle the earth, particularly over the oceans, in both the Northern Hemisphere and Southern Hemisphere, with a number of centers of maximums about 30 to 35 degrees from the Equator.

SEASONAL VARIATIONS

There are also regions where the pressure is predominantly low or high at certain seasons, but not throughout the year.

In the vicinity of Iceland, pressure is low most of the time. The water surface is warmer than the icecaps of Greenland and Iceland. The Icelandic low is most intense in winter, when the greatest temperature differences occur, but it persists with less intensity through the summer. The Aleutian low is most pronounced when the neighboring areas of Alaska and Siberia are snow covered and colder than the adjacent ocean.

These lows are not a continuation of one and the same cyclone. They are regions of low pressure, where lows frequently form or arrive from other regions to remain stationary or move sluggishly for a time, after which they pass on or die out and are replaced by others. Occasionally these regions of low pressure are invaded by traveling high-pressure systems.

There is a semipermanent high-pressure center over the Pacific to the westward of California. Another overlies the Atlantic, near the Azores and off the coast of Africa. Pressure is also high, but less persistently so, westward of the Azores to the vicinity of Bermuda. (Both are best developed in the summer season.)

The largest individual circulation cells in the Northern Hemisphere are the Asiatic high in winter and the Asiatic low in summer. In winter, the Asiatic continent is a region of strong cooling and therefore is dominated by a large high-pressure cell. In summer, strong heating is present and the high-pressure cell becomes a large low-pressure cell. This seasonal change in pressure cells gives rise to the monsoonal flow of the India-S.E. ASIA area.

Another cell which some consider to be a center of action is the polar high. However, recent explorations into both the Arctic and the Antarctic have revealed considerable variations in pressure in these regions and the presence of many traveling disturbances in summer. The Greenland high, for example, due to the Greenland icecap, is a persistent feature, but it is not a well-defined high during all seasons of the year. It often appears to be an extension of the polar high, or vice versa.

Other continental regions show seasonal variations from low pressure in summer, but are generally of small size, and their location is variable. Therefore, they are not considered to be centers of action.

The annual average pressure distribution chart for the year, (fig. 4-5) reveals several important characteristics.

First, along the Equator there is a belt of relatively low pressure encircling the globe with barometric pressure of about 1,012 millibars. Second, on either side of this belt of low pressure is a belt of high pressure. That in the Northern Hemisphere lies mostly between latitudes 30° and 40° N with three well-defined centers of maximum pressure—one over the eastern Pacific, the second over the Azores, and the third over Siberia, all about 1,020 millibars. The belt of high pressure in the Southern Hemisphere roughly follows parallel 30° S. It also has three centers of maximum pressure—one in the eastern Pacific, the second in the eastern Atlantic, and the third in the Indian Ocean, again about 1020 millibars.

A third characteristic to be noted from this chart is that beyond the belt of high pressure in either hemisphere the pressure diminishes toward the poles. In the Southern Hemisphere the decrease in pressure toward the South Pole is regular and very marked. The pressure decreases from an average somewhat above 1,016 millibars along latitude 35° S to an average of 992

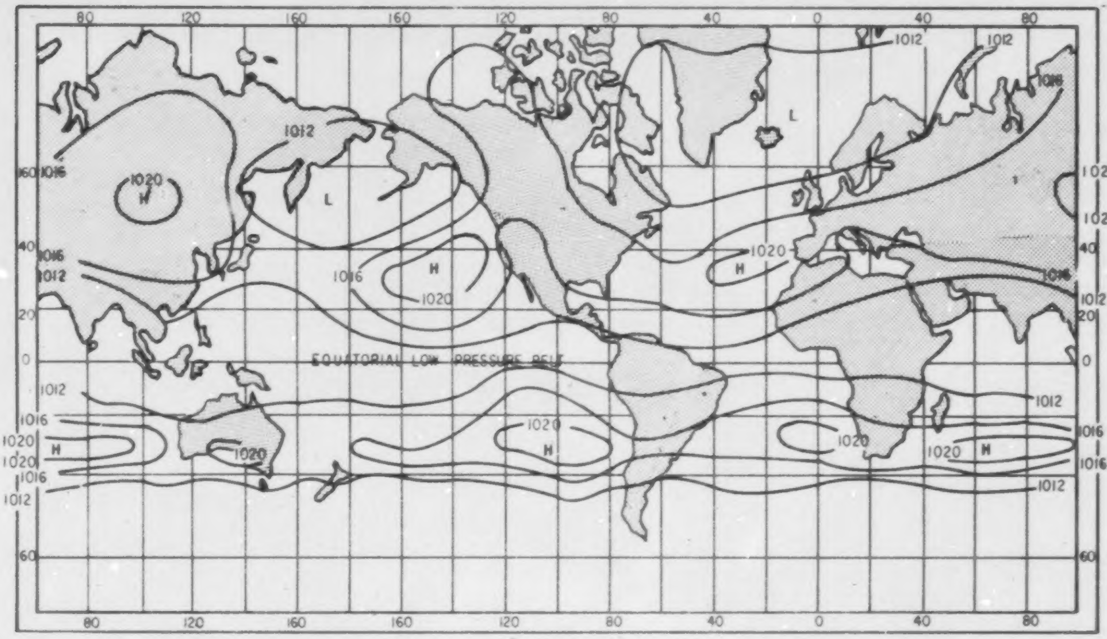

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Figure 4-5.—Average annual pressure distribution chart.

millibars along latitude 60° S. In the Northern Hemisphere, however, the decrease in pressure toward the North Pole is less regular and not so great.

While the pressure belts which stand out on the average annual pressure distribution chart represent average pressure distribution for the year, these belts are rarely continuous on any given day. They are usually broken up into detached areas of high or low pressure by the secondary circulation of the atmosphere. In either hemisphere, moreover, the pressure over the land during the winter season is decidedly above the annual average, during the summer season decidedly below it, the extreme variations occurring in the case of continental Asia where the mean monthly pressure ranges from about 1,033 millibars during January to about 999 millibars during July. Over the northern oceans, on the other hand, conditions are reversed, the summer pressure there being somewhat higher. Thus in January the Icelandic and Aleutian lows intensify to a depth of about 999 millibars, while in July these lows fill up and are almost obliterated.

The subtropical highs reach their greatest intensity during the summer and the area over which they extend is much greater and the pressure higher in summer. In winter these subtropical highs are found at lower latitudes and migrate poleward with the onset of the summer season and equatorward with the winter season.

The polar high in winter is not a cell centered directly over the North Pole, but it appears to be an extension of the Asiatic high and often appears as a wedge extending from the Asiatic Continent. The cell is displaced toward the area of coldest temperatures, the Asiatic Continent. In summer, this high appears as an extension of the Pacific high and is again displaced toward the area of coolest temperature, which in this case is the extensive water area of the Pacific.

In winter over North America, the most significant feature is the domination by high-pressure cells. These cells are also due to cooling but are not as intense as the Asiatic cells.

In summer, the most significant feature is the so-called heat low over the southwestern part of the continent which is caused by extreme heating in this region.

TRAVELING DISTURBANCES

These disturbances are part of the secondary circulation, but do not show up on mean pressure charts. They are migratory and include such systems as the lows which form along the polar front and the migratory highs which accompany polar outbreaks into the middle latitudes. These traveling disturbances play an important role in the day-to-day changes in the weather.

TERTIARY CIRCULATIONS

The preceding sections deal with the general circulation and the secondary circulation and the main forces causing windflows, and the wind patterns found in high- and low-pressure areas. However, the concepts of the large-scale movements of the wind do not take into consideration the effects of local conditions which frequently cause drastic modifications in wind direction and speed close to the earth's surface. Thus, from the point of view of the meteorologist a knowledge of the factors which affect local winds is desirable.

Tertiary circulations are localized circulations directly attributable to one of the following causes or a combination of them: local cooling, local heating, adjacent heating or cooling, and induction (dynamics).

MONSOON WINDS

A discussion on monsoon winds was presented in chapter 5, AG 3 & 2, as well as in chapter 2 of this training manual. Therefore the following paragraph is confined to a brief mention of the general type of weather which may be expected where monsoon conditions are in existence.

The northeast (winter) monsoon blows in the South China Sea from October to April. It is marked by dry or fair weather with the exception that the Gulf of Tonkin and the coast of North Vietnam experience precipitation and low visibilities due to fog as a result of periodic interruptions of the monsoon flow as the outbreaks of polar air are modified by their trajectories over water areas to the northeast. It has a steadiness similar to that of the trade winds and often attains the force of a near gale. The southwest or summer monsoon occurs from May to September. It breaks with severity on some coasts, accompanied by heavy squalls and thunderstorms. As the season advances and the southwest monsoon becomes established, squalls and rain become less frequent. In some places it blows as a light breeze, unsteady in direction, while in other places it prevails with fresh speeds throughout the season, being infrequently interrupted by calms or by winds from other directions.

LAND AND SEA BREEZES

Corresponding with the seasonal contrast of temperature and pressure over land and water (such as the monsoon effect) there is a diurnal contrast exercising a similar, though more local, effect. An example of this type of influence on the existing wind flow is the "land breeze," or its reverse condition commonly referred to as the "sea breeze." These two diurnal breezes were presented previously in chapter 5 of AG 3 & 2, therefore, only brief mention of them is made at this time.

The sea breeze usually begins in the morning hours, from 9 to 11 o'clock local time, and is usually only a few hundred feet thick. By midafternoon, when it has reached its maximum speed, it can extend upward to 3,000 feet in moderately warm climate and 4,500 feet in tropical regions and to as much as 30 miles both on shore and offshore. Also, by midafternoon it may be strong enough to be influenced by the Coriolis force, causing it to flow at an angle to the shore.

The land breezes, when compared to the sea breeze, are less extensive and not as strong. Land breezes are at maximum developement late at night, in late fall, and early winter.

In the Tropics, the land and sea breezes are repeated day after day with great regularity. In high latitudes the land and sea breezes are often masked by winds of synoptic features.

The occurrence or nonoccurrence of the sea breeze and the return land breeze flow is of critical importance in the evaluation of the effects of the release of airborne pollutants on individual harbors. The lapse rate within the sea breeze layer is approximately adiabatic in the temperate latitudes. The moderate to strong wind speeds combined with this lapse rate would result in excellent dispersal of material throughout the sea breeze layer. However, the top of the sea breeze layer is likely to be marked by an inversion or at least a decrease in lapse rate and wind velocity so that a fairly effective upper barrier to the diffusion of material would exist. In addition, the sea breeze decreases steadily in depth with its continued penetration inland, so that the top of the leading edge may be only a few tens of feet above the ground and thus confine material very effectively to a shallow layer.

The land breeze is quite stable, the wind speeds are light, and diffusion is at a minimum. On accasion the land breeze is insufficient to carry the material more than a mile or so offshore and, with the reestablishment of the sea breeze, this material may be recirculated onto the land.

WINDS DUE TO LOCAL COOLING

Drainage Winds

Drainage winds, also called mountain or gravity winds, are caused by cooling of air along the slopes of a mountain. Consequently, the air becomes heavy and flows downhill producing the MOUNTAIN BREEZE.

These mountain breezes, which were previously discussed in chapter 5 of AG 3 & 2, may reach several hundred feet in depth and in extreme cases may attain speeds of 50 knots or more. The downhill motion of these and other breezes with similar motion is termed "katabatic." They may be due to local terrain configuration or large scale causes such as air mass characteristics. Although drainage winds and the glacier winds discussed in the following paragraphs are cold katabatic winds, in some instances katabatic winds may be warm. The foehn winds described later in this section are an example of warm katabatic winds. The term used to describe the opposite or upslope motion of wind is "anabatic." These winds are also presented later in this section.

Glacier Winds

These winds are caused by air coming in contact with ice or some other cold surface, as explained in chapter 5 of AG 3 & 2. This will cause faster cooling than is normal and the air will be set in motion by a strong pressure gradient or density difference. Also, if this wind is funneled through a pass or valley, it may be very strong. This type wind may form during the day or night, while the mountain breezes occur only at night due to radiational cooling. The glacier wind is most common during the winter when more snow and ice are present.

Glacier winds, as mentioned previously, are cold katabatic winds descending from glacier regions and are localized in nature.

Figure 4-6 shows how light winds can often become much stronger when they are forced to converge and funnel through a narrow mountain pass. The Santa Ana wind of southern California (described later in this chapter) is due in part to this type of phenomenon.

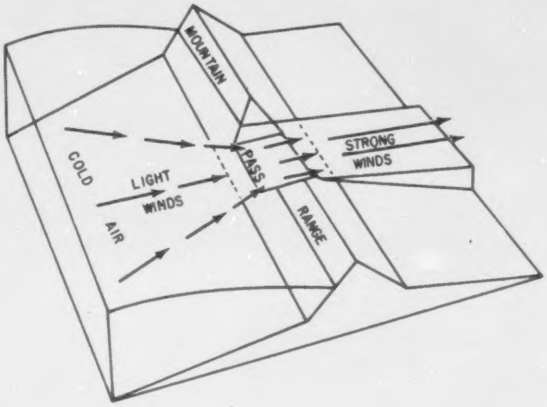

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Figure 4-6.–Strong wind produced by funneling.

WINDS DUE TO LOCAL HEATING

There are three types of tertiary circulation caused by local heating: valley breezes, dry thermals, and wet thermals.

Valley Breeze

Valley breezes or anabatic winds are produced when the daytime air near a mountain is heated by contact while air at some distance from the mountain surface is not affected. This produces light air near the slope which tends to move upward to high levels and thereby produces the valley wind. They are generally restricted to slopes facing the south or the more direct rays of the sun and are more pronounced in southern latitudes. They are diurnally strongest during late afternoon and are seasonally strongest in summer. They are deeper and stronger than mountain or katabatic winds.

It is conceivable that this type of circulation could be found in an area where there would be very little net outflow of air from the valley system, If this were the case, a pollutant contained within the valley air mass would be mixed by this circulation until most of it was diffused throughout the depth of circulation. As a result, a much heavier ground concentration of pollutants might exist in this region than would be expected with a more rapid change of air mass within the valley system.

Dry Thermals

Dry thermals are small local vertical circulations caused by the uneven heating of the air due to the fact that some kinds of surfaces are more effective than others in heating the air directly above them. Plowed ground, sand, rocks and barren land give off a great deal of heat, whereas water and vegetation do not. When the dry thermal tops do not reach the condensation level, they are commonly referred to as dust devils. They are best developed on clear days with light winds and high temperatures. Dust devils are most frequently seen over desert areas and can reach as much as 5,000 feet under extremely dry air conditions. When occurring over water areas, dust devils are called waterspouts.

Wet Thermals

Cumulus type clouds occur when rising wet thermals reach their condensation level. They are topped with a cumulus type cloud with the base at the condensation level. The tops are low under stable conditions, but may develop into cumulonimbus under unstable conditions.

INDUCED OR DYNAMIC TERTIARY CIRCULATIONS

Induced or dynamic tertiary circulations are of three types: eddies, foehn winds, and jet winds.

Eddy Winds

Eddies derive their energy from the windflow over topographical barriers or obstructions. They are on a much larger scale when the obstacle is a mountain barrier. The eddies are generally found on the lee side of mountains, but with low wind speeds, stationary eddies or rotating pockets of air are produced and remain on both the windward and leeward sides of obstructions, such as buildings over which the air may be passing. (See fig. 4-7.)

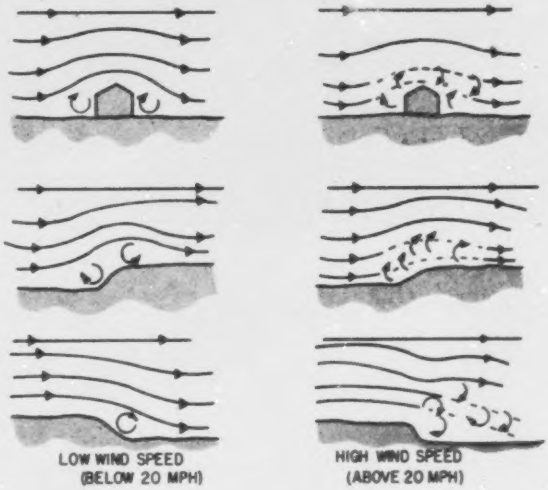
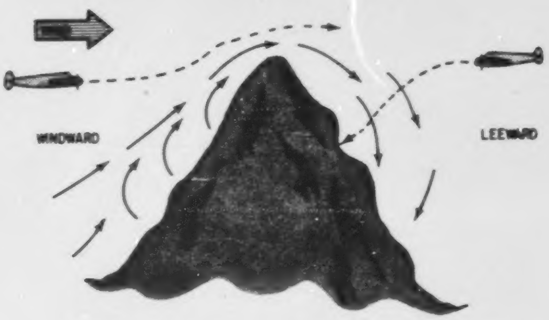


Figure 4-7.—Eddy currents formed when wind flows over uneven ground or obstructions.

When the wind speeds exceed about 20 miles per hour, the flow may be broken up in irregular eddies which are carried along with a wind some distance downsteam from the obstruction. These eddies may cause extreme and irregular variations in the wind and may disturb the landing area sufficiently to be a hazard. It is important to be alert for these turbulent eddies when a landing field is located near large hangars or other buildings.

A similar and much disturbed wind condition occurs when the wind blows over large obstructions such as mountain ridges. In such cases the wind blowing up the slope on the windward side is usually relatively smooth. However, on the leeward side the wind spills rapidly down the slope, setting up strong downdrafts and causing the air to be very turbulent. This condition is illustrated in figure 4-8.



AG.418

Figure 4-8.—Effect of windflow over mountains.

The situation can be compared to water flowing down a rough streambed. These downdrafts can be very violent, and they have been the cause of a number of aircraft accidents, causing the aircraft to crash into the sides of mountains. This effect is also noticeable in the case of hills and bluffs, but is not as pronounced.

Eddy winds may also result in a change of location of pollutant material within an air mass, or they may result in the mixing of clean air with the polluting material. The latter decreases the concentration of material per unit volume and is of primary concern. A puff of material, if actually injected into a volume of air, will seem to grow larger. This is due almost entirely to eddy motions. Molecular motion also produces diffusion, but compared to eddy diffusion, molecular diffusion is relatively insignificant.

Foehn Winds

When air flows downhill from a high elevation, its temperature is raised by adiabatic compression. Foehn winds are katabatic winds caused by adiabatic heating of air as it descends on the lee sides of mountains. The rise in temperature results in a lowering of the relative humidity. They arrive at the bottom of the lee side of the mountain causing temperature rises as much as 50°F.

Foehn winds occur quite frequently in the western Mountain States. In Montana and Wyoming the chinook is a well known phenomenon, and in Southern California the Santa Ana is known particularly for its high-speed winds.

Generally speaking, when the Santa Ana blows through the Santa Ana Canyon, a similar wind simultaneously affects the entire southern California area. Thus, when meteorological conditions are favorable this dry northeast wind will blow through the many passes and canyons, over all the mountainous area, including the highest peaks, and quite often at exposed places along the entire coast from Santa Maria to San Diego. Therefore, the term Santa Ana refers to the general condition of a dry northeast wind over southern California.

Although these winds may on occasion reach destructive velocities, one beneficial aspect to consider is that when they coincide with the release of airborne pollutants over an area, the material would be quickly dispersed and carried away from the area affected.

A complete study and discussion of the Santa Ana may be found in Climatology and Low-Level Air Pollution Potential From Ships in San Diego Harbor, NWRF 39-0462-056.

Jet Winds

Jet winds, also known as mountain gap or canyon winds, are extensive squall type winds, whose speed is increased through a channeling effect. See figure 4-6 for the effect of mountain gaps.

LARGE-SCALE VERTICAL WAVES (MOUNTAIN WAVES)

This type of phenomenon occurs on the lee side of topographical barriers and occurs when the windflow is strong, usually 25 knots or more, and the flow is roughly perpendicular to the mountain range. The structure of the barrier and the strength of the wind determines the amplitude and the type of the wave. The characteristics of a typical mountain wave are shown in figure 4-9.

Figure 4-9 shows the cloud formations normally found with wave development and illustrates schematically the airflow in a similar situation. From the illustration in figure 4-9 it can be seen that the air flows fairly smoothly with a lifting component as it moves along the windward side of the mountain. The wind speed gradually increases reaching a maximum near the summit. On passing the crest the flow breaks down into a much more complicated pattern with downdrafts predominating. An indication of the possible intensities can be gained from verified records of sustained downdrafts (and also updrafts) of at least 3,000 feet per minute with other reports showing drafts well in excess of this figure. Turbulence in varying degrees can be expected as indicated in figure 4-9 and is most likely to be particularly severe in the lower levels. Proceeding downwind, some 5 to 10 miles from the summit, the airflow begins to ascend as part of a definite wave pattern. Additional waves, generally less intense than the primary wave, may form downwind (in some cases six or more have been reported). These are not unlike the series of ripples that form downstream from a submerged rock in a swiftly flowing river. The distance between successive waves usually ranges from 2 to 10 miles, depending largely on the existing wind speed and the atmospheric stability, but wave lengths up to 20 miles have been reported.

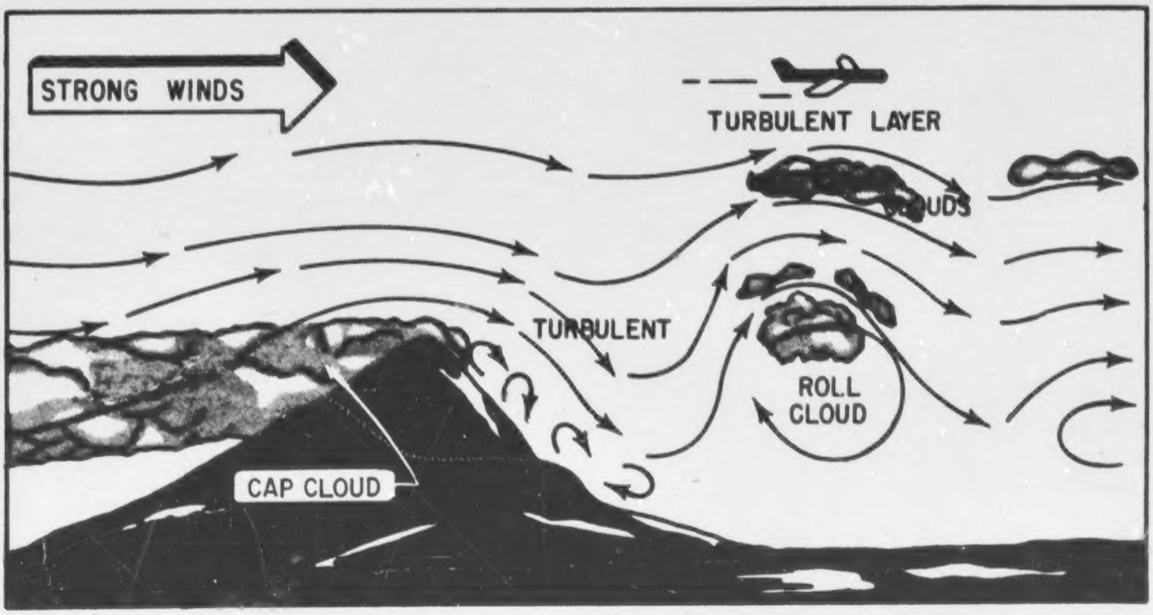


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Figure 4-9.—Schematic diagram showing airflow and clouds in a mountain wave.

From the meteorologist's standpoint it is important to know how to identify a wave situation, and having identified it to advise the pilot how to plan his flight so as to avoid the wave hazards. Characteristic cloud forms peculiar to wave action provide the best means of visual identification. The lenticular (lens shaped) clouds in the upper right of figure 4-9 are smooth in contour. These clouds may occur singly or in layers at heights usually above 20,000 feet, and may be quite ragged when the airflow at that level is turbulent. The roll cloud forms at a lower level, generally near the height of the mountain ridge, and can be seen extending across the center of figure 4-9. The cap cloud shown partially covering the mountain slope to the left of figure 4-9 must always be avoided in flight because of the turbulence, concealed moutain peaks, and on the lee side strong downdrafts. The lenticulars, like the roll clouds and cap clouds, are stationary, constantly forming on the windward side and dissipating on the lee side of the wave. The cloud forms themselves are a good guide to the degree of turbulence with generally smooth airflow in and near smooth clouds, and turbulent conditions if these clouds appear ragged or irregular.

While clouds are generally present to forewarn the presence of wave activity, it is possible for wave action to take place when the air is too dry to form clouds. This makes the problem of identifying and forecasting more difficult.

VERTICAL EXTENSION OF HIGH- AND LOW-PRESSURE CELLS

In order to better understand the nature of the pressure centers of the secondary circulation which was covered previously in this chapter, it is necessary to consider them from a three-dimensional standpoint, not only length and width, but vertically as well. With the aid of upper air charts, you will be able to see the three dimensions of these pressure systems, as well as the circulation patterns of the secondary circulation as established at higher levels in the troposphere and lower stratosphere. With the knowledge gained through this study you will be better able to understand the mechanics of the jetstream and some of the forecasting rules which are covered in later chapters.

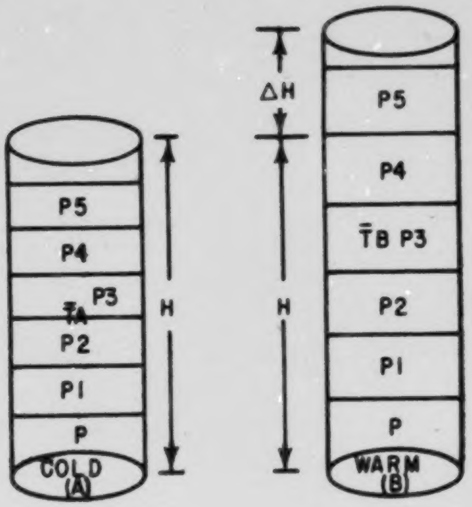
MEAN TEMPERATURE AND VERTICAL SPACING OF ISOBARS

In chapter 3 of this training manual, Atmospheric Physics, and application of the gas laws shows that volume is directly proportional to temperature. Stated another way, we might say that the thickness of a layer between two isobaric surfaces is directly proportional to the mean virtual temperature of the layer. Thus, thickness lines are also isotherms of mean virtual temperature. The higher the mean virtual temperature, the thicker the layer, or vice versa. The thickness between layers is currently being expressed in geopotential meters. The shift in location, as well as the change of shape and intensity upward of atmospheric pressure systems, is dependent on the temperature distribution.

An example of the foregoing is shown by the fact that if two columns of air are placed side by side, one cold and one warm, the constant pressure surfaces in the cold column will be closer together than those in the warm column of air. Figure 4-10 illustrates an increase in thickness between two given pressure surfaces for an increase in mean virtual temperature. Also, note the increase in the distance between the constant pressure surfaces; P, P1, etc., from A to B.

The thickness between two pressure surfaces can be derived by integrating the hydrostatic equation, or by means of the hypsometric equation. (See ch. 3.) Thickness may also be constructed from known variables by the use of tables, graphs, etc.

Within the troposphere, the horizontal temperature gradient is directed poleward from the lower few thousand feet of the troposphere up to the level of the tropopause. The maximum gradient is in middle latitudes and reflects a seasonal shift southward in winter and northward in summer. This maximum gradient also has its highest numerical value in winter. The minimum temperature occurs at the tropopause. The subarctic tropopause is much lower in winter than in summer and is always lower than the subtropical tropopause.



ΔH IS THE INCREASE IN THICKNESS BETWEEN TWO GIVEN PRESSURE SURFACES FOR AN INCREASE IN MEAN VIRTUAL TEMPERATURE FROM T̄A TO T̄B. T̄B IS A HIGHER MEAN VIRTUAL TEMPERATURE THAN T̄A.

AG.420

Figure 4-10.—Thickness of two strata as a function of mean virtual temperature.

Within the lower stratosphere during summer, the temperature gradient is directed from the poles to the Equator. During winter, the temperature gradient is directed from approximately 55° lat to the Equator, and from 55° lat to the poles. This gives rise to a wind condition called the polar vortex. Within the upper stratosphere during summer, the temperature gradient is still directed from the poles to the Equator, though much weaker than in the lower stratosphere. During winter, the gradient is directed from the Equator to the poles in the upper stratosphere.

In the mesophere in summer, the temperature gradient is directed away from the Equator to the poles, while in winter it is from the poles to the Equator. In the thermosphere, the distribution is unconfirmed, though it is believed that during summer the gradient is from the poles to the Equator and vice versa in winter.

VERTICAL STRUCTURE OF HIGH-PRESSURE CELLS

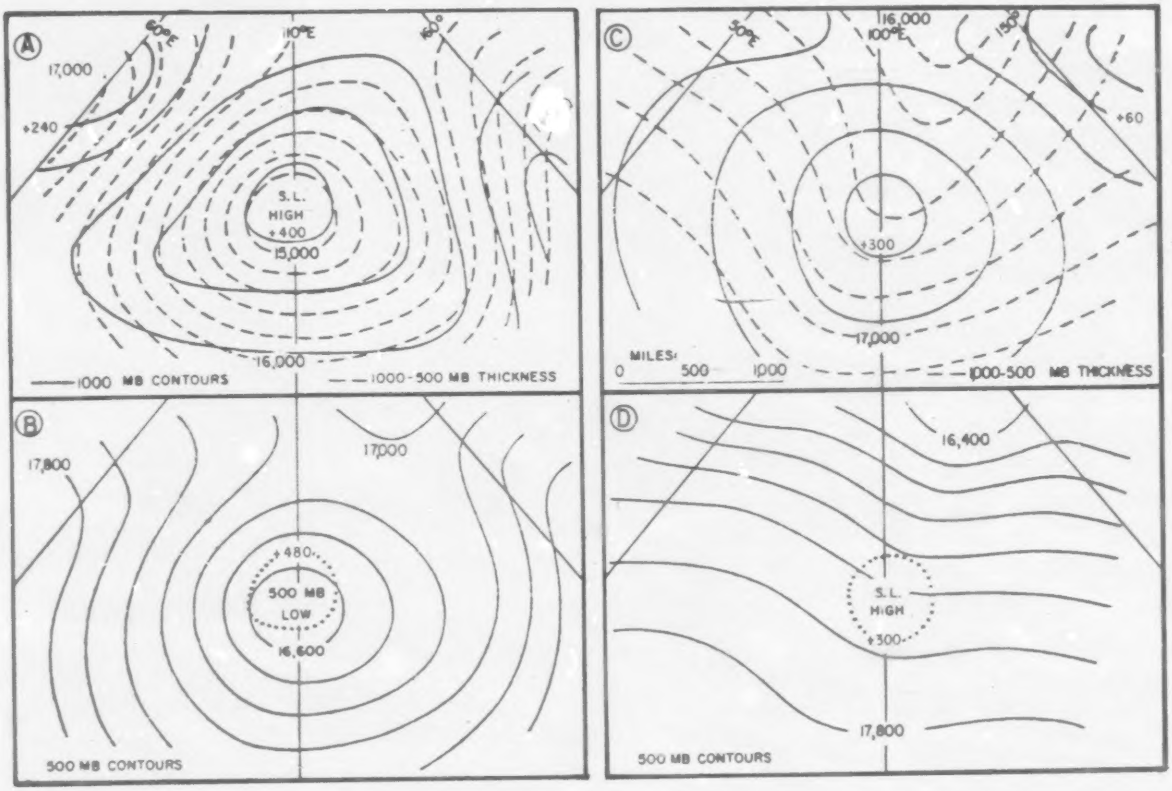
The topographic features which indicate the circulation patterns at 500 milibars in the atmosphere correspond in general to those at lower and higher levels. However, they may experience a shift in location as well as a change in intensity and shape. For example, a closed high on a surface synoptic chart may be reflected by a ridge aloft. In addition, upper air circulation patterns may take on a wavelike structure in contrast to the alternate closed lows, or closed high patterns at the surface level. The smoothing of the circulation pattern aloft is typical of atmospheric flow patterns. The vertical isobaric spacing of the following types of highs is illustrated in chapter 21, AG 3 & 2, NavTra 10363-D.

Cold Core Highs

The cold core high is so characterized due to its low tropospheric coldness and exists principally because of the weight of the cold air near the surface. The cold high disappears rapidly with height. If it becomes subjected to warming from below and subsidence aloft, as it moves southward from its source and spreads out, it diminishes rapidly in intensity with time (unless some dynamic effect sets in aloft over the high to compensate for the warming). Since these highs decrease in intensity with height, thicknesses are relatively low. If the thickness pattern is sufficiently strong, the wind circulation around the center can reverse its direction at high levels.

Usually the temperature field around a cold-core anticyclone is asymmetrical. The center of low thickness will be found displaced from the 1,000-mb high toward the region of lowest temperatures. Thicknesses are relatively greater where middle cloud and cirrus are present or precipitation is falling from layer clouds (except stratocumulus).

Figure 4-11 (A) shows an extremely cold and intense sea level anticyclone and corresponding thickness pattern for the 1,000 to 500-mb layer, the latter for a very deep cold pool (low thickness) centered directly over the 1,000-mb high.


A
60°E
110°E
160°
17,000
+240
S.L.
HIGH
+400
15,000
16,000
1000 MB CONTOURS
1000-500 MB THICKNESS
C
50°E
16,000
100°E
150°
+60
+300
17,000
MILES:
0
500
1000
1000-500 MB THICKNESS
B
17,800
17,000
+480
500 MB
LOW
16,600
500 MB CONTOURS
D
16,400
S.L.
HIGH
+300
17,800
500 MB CONTOURS

AG.421

Figure 4-11.—Two examples of cold core anticyclones in relation to 1,000- and 500-mb contours and 1,000- to 500-mb thickness.

Figure 4-11 (B) shows the associated 500-mb pattern, containing a closed 500-mb low directly over the 1,000-mb high. This model characterizes extreme cases over continental areas of middle and high latitudes. It is based on actual observations of the Siberian winter anticyclone.

Less intense Siberian winter anticyclones and most North American winter anticyclones of the sub-Arctic regions are characterized by the models illustrated in figures 4-11 (C) and (D). Figure 4-11 (C) contains 1,000-mb contours and 1,000- to 500-mb thickness lines. A cold trough (low thickness) lies over the sea level anticyclone, with the lowest thickness values to be found in polar areas well to the north. Figure 4-11 (D) shows the corresponding 500-mb contour lines, which form a weak trough whose axis lies over the center of the sea level anticyclone.

An illustration of a cold core high (and warm core low) is shown in figure 4-12.

Warm Core Highs

The warm high is so characterized because throughout the troposphere the warmest air is at the center of the high. This high has an anticyclonic circulation at all levels; hence the term "high level anticyclone." In the stratosphere this type of high has a cold core associated with the center of the anticyclonic circulation.

In the warm core high the thickness values over the center are higher than along the periphery and the thickness center is displaced toward the warmest air in cases of an asymmetric temperature distribution. These highs are characteristically found in the form of large warm, oceanic highs. The subtropical highs are good examples of this type of high. In middle and high latitudes it is found in the form of warm blocking highs.

As with other models, weather manifestations in a warm high provide clues to the thickness distribution. Convective activity usually is associated with somewhat lower thickness values; thus the thickness center (high) is displaced to the side of the 1,000-mb anticyclone away from the areas of convection.

VERTICAL STRUCTURE OF LOW-PRESSURE CELLS

Warm Core Lows

A warm core low is similar to a cold core anticyclone in that it decreases in intensity with height. This type low is characterized by virtue of its low tropospheric warmth. In low tropospheric thickness charts, a warm core or tongue of maximum thickness is found. It is apparent that if the thickness pattern is pronounced enough, the sum of the 1,000-mb height at the low center and the thickness over the low can produce an anticyclone at the upper surface of the layer. They rarely affect the stratospheric circulation. The warm low is frequently stationary, such as the heat low over the southwestern United States in the summer, and is a result of strong heating in a region usually insulated from intrusions of cold air which would tend to fill it or cause it to move. The warm low is also found in its moving form as a stable wave moving along a frontal surface. It moves by virtue of the low level advection in advance of it and the cold advection to its rear. This type of wave cyclone lacks any dynamic mechanism aloft to cause it to deepen and occlude. It usually fills as it moves eastward, and the frontal wave flattens out. There is no warm low aloft in the troposphere. The tropical cyclone is believed to be a warm low; its intensity diminishes with height.

Figure 4-12 shows an idealized warm core cyclone with symmetrical temperature distribution.

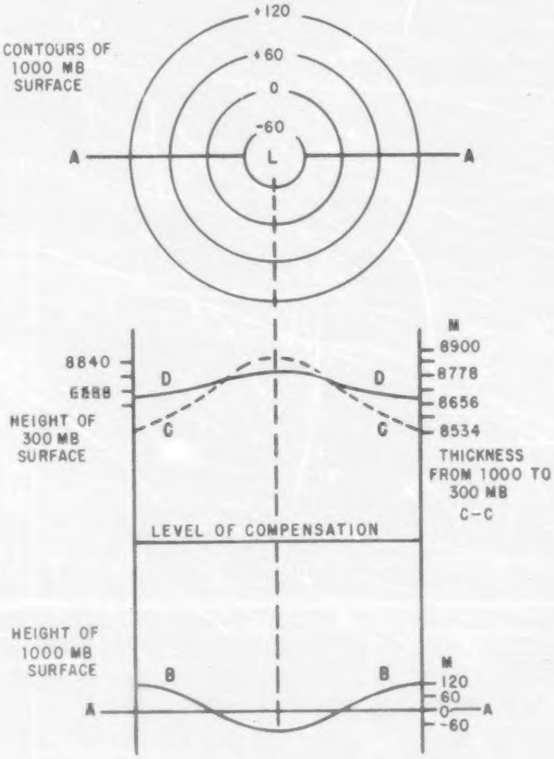


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Figure 4-12.—Idealized warm core cyclone with symmetrical temperature distribution.

This figure shows how the thickness values indicate the presence of an anticyclone aloft over the low. The upper diagram represents idealized contours of the 1,000-mb surface. Curve BB shows the profile of the 1,000-mb surface along lien AA, curve CC shows the 1,000- to 300-mb thickness profile along the same line, and curve DD is the sum of BB and CC, or the profile of the upper pressure surface along AA. In this idealized example the thickness (temperature) and coutour fields are mutually symmetrical.

In general, however, the temperature field is quite asymmetrical around a warm core cyclone. Usually the southward moving air in the rear of the depression will not be as warm as that moving northward in advance of it. Here the maximum thickness values will be found on the warmest side of the cyclone, whose axis will slope upward away from the thickest region of the layer. An examination of the thickness field associated with this warm core low shows the largest thicknesses are found in the warm sector just to the east of the wave crest. The thickness' circulation pattern with the warm core low is opposite the circulation pattern at the surface, and the low decreases in intensity with increasing height. The winds, as might be expected, in this region shift from northeasterly at the surface to southwesterly aloft. The surface low is replaced by a ridge aloft.

Weather and cloudiness found in warm core cyclones can yield clues to the thickness. Where evidence of convective activity, such as showers, is greatest, the thickness values are relatively low although they are in a region of generally high values. Where the weather is fine and clear, except for perhaps a few cirrus, the thickness values will be relatively high.

Cold Core Lows

The cold core low in its purest form contains the coldest air at its center throughout the troposphere; that is, going radially outward in any direction, at any level in the troposphere warmer air is encountered. The cold core lows increase in intensity with height. Relative minimums in thickness values, called cold pools are found in such cyclones. When the temperature distribution is asymmetric, the thickness center will lie in the direction of the lowest temperatures. When the temperature distribution is symmetric, the axis of the low is nearly vertical. Tropospheric thickness charts show closed minimum thickness lines symmetrical with the pressure center. In other words, in the cold low, the lowest temperatures coincide with the lowest pressures.

In the lower stratosphere, there is a warm core vertically over the low center. Here the highest temperatures are associated with the lowest pressures.

The cold low has a more intense circulation aloft from 850 to 400 millibars than at the surface. Some cold lows show but very slight evidence in the surface pressure field that an intense circulation exists aloft. The cyclonic circulation aloft is usually reflected on the surface in an abnormally low daily mean temperature and in precipitation and unstable hydrometeors.

At high latitudes the cold pools and their associated upper air lows show some tendency for location in the northern Pacific and Atlantic Oceans where, statistically, they contribute to the formation of the Aleutian and Greenland lows. At lower latitudes the assoicated upper air cyclones are frequently called CUTOFF LOWS, which usually occur when blocking highs exist far to the north.

Figure 4-13 illustrates the displacement of a cold core low southwestward aloft where the lowest thickness values are displaced toward the region of lowest temperatures. Also note the position of the 500-mb low.

The thickness circulation pattern over this low has the same direction as the surface circulation pattern. This indicates an increase in winds aloft. At 500 millibars the winds are seen to be considerably stronger than at the surface and the wind direction does not shift appreciably with height.

DYNAMIC LOW

The dynamic low is a combinaion of the warm surface low and a cold upper low or trough, or a warm surface low in combination with a dynamic mechanism aloft for producing a cold upper low or trough. It has an axis which slopes toward the coldest tropospheric air. In the final stage, after occlusion of the surface warm low is complete, the dynamic low becomes a cold low with the axis of the low becoming practically vertical.

DYNAMIC HIGH

The dynamic high is a combination of a surface cold high and an upper-level warm high or well-developed ridge, or a combination of a surface cold high with a dynamic mechanism aloft for producing high-level anticyclogenesis. Dynamic highs have axes which slope toward the warmest tropospheric air. In the final stages of

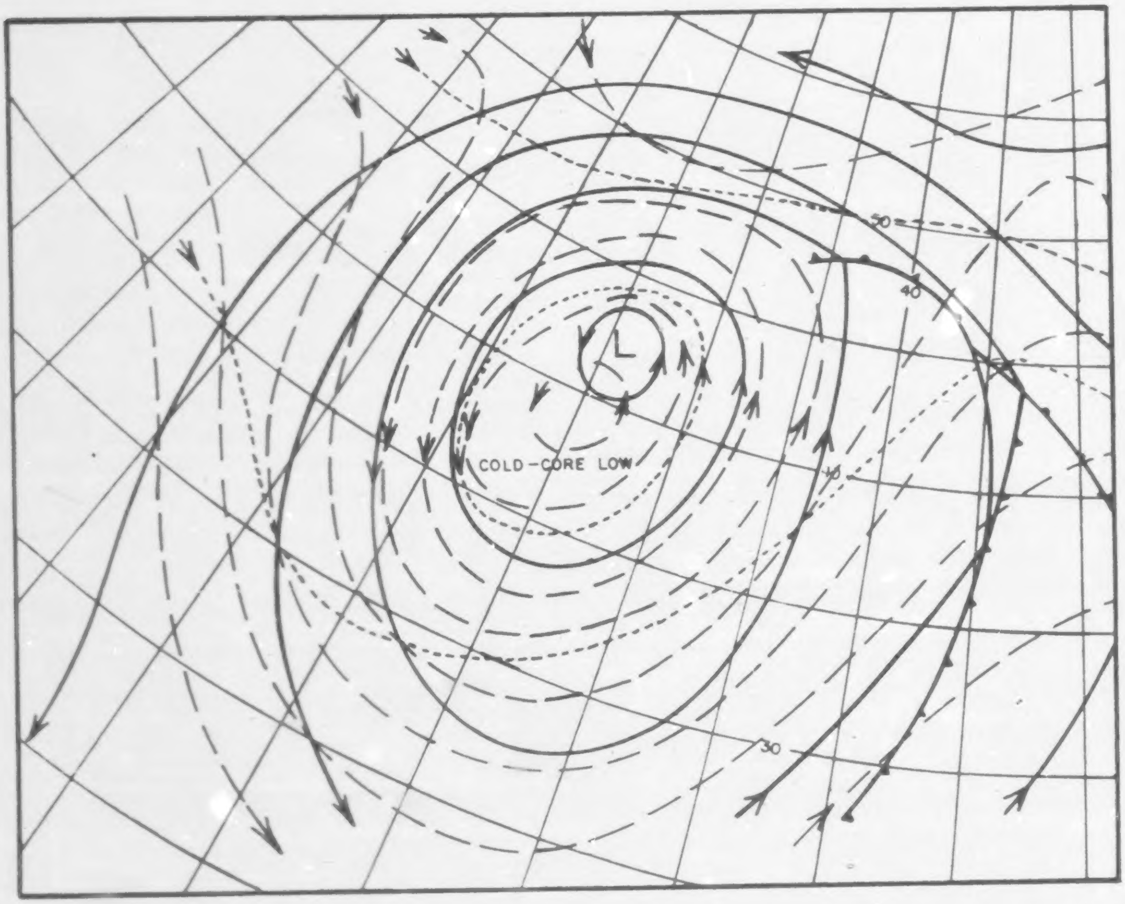

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Figure 4-13.—Illustration of a cold core low showing 1,000-mb contours (solid lines), 500-mb contours (dashed lines), and thickness pattern (dotted lines).

warming of the cold surface high, the dynamic high becomes a warm high with its axis practically vertical.

SUMMARY

A warm core high is accompanied by a high cold tropopause. Since the pressure surfaces are spaced far apart, the tropopause is not reached until great heights. The temperature continues to decrease with elevation and is very cold by the time the tropopause is reached. The subtropical highs are good examples of this type of high. Therefore, anticyclones found in tropical air are always warm cored. Anticyclones found in Arctic air are always cold cored, while anticyclones in polar air may be warm or cold cored.

A cold core low is accompanied by a low warm tropopause. Since the pressure surfaces are close together, the tropopause is reached at low altitudes where the temperature is relatively warm. Good examples of cold core lows are the Aleutian and Icelandic lows. Occluded cyclones will generally be cold cored because of the polar or Arctic air that has closed in on them.

Warm core lows decrease in intensity with height or completely disappear and are for the most part replaced by anticyclones aloft. The heat lows of the southwestern United States, and over Asia and Africa are good examples of warm core lows. Newly formed waves will generally be warm cored because of the wide-open warm sector.

Systems which retain their closed circulations to appreciable altitudes are called dynamic lows or highs.

USE OF CONSTANT PRESSURE CHARTS

The analysis of and a general description of the types of constant pressure charts is presented in chapter 21, AG 3 & 2, NavTra 10363-D. Therefore this section will be confined to a brief description of the application of the various constant pressure charts.

THE 1,000-MB CHART

This chart indicates the height of the 1,000-mb pressure surface above or below sea level. When the 1,000-mb surface lies below sea level, it is indicated by negative height values. In mountainous areas caution must be observed in using 1,000-mb height values.

The chart is normally constructed from the surface chart by assuming that 7 1/2 millibars equal 60 meters (200 feet) for temperatures between 0° and 20° C. For temperatures below 0° C 8 1/2 to 9 millibars equal 60 meters, and for temperatures above 20° C, 6 to 6 1/2 millibars equal 60 meters. These are a valid approximation of the 1,000-mb heights. Heights from radiosonde soundings should be used whenever available.

The principal use of the 1,000-mb chart is made in constructing space differential (thickness) charts. It serves as the base level for the 1,000- to 700-mb and the 1,000- to 500-mb differential charts.

THE 850-MB CHART

The approximate height of the 850-mb level according to U.S. Standard Atmosphere is 1,460 meters (4,780 feet). The principal use of this chart is to determine the representativeness of surface winds and temperatures, to determine depth of moisture patterns in winter, and to replace the surface chart in mountainous and plateau areas where the mean elevation is around 5,000 feet. Both temperature (frontal) analysis and moisture analysis should be carried out on this chart, and its analysis should always be made in close conjunction with the surface chart whenever possible.

A complete and careful isotherm analysis made at this level in conjunction with wind and pressure analysis will lead to the correct placement of fronts at this level and by implication location of the surface front. A thumb rule to guide Aerographer's Mates in location of most fronts is to look for the 850-mb warm front roughly 2 1/2° to 3° latitude ahead of the surface front and cold fronts from 3/4° to 2° latitude behind the surface front.

The interval is the same as the 1,000-mb chart–60 meters. Radiosonde and moutain station reports of the height of the 850-mb level should be used in plotting and analyzing this chart. The 850-mb isotherms serve as a good indication of the 1,000- to 700-mb thickness patterns.

THE 700-MB CHART

The approximate height of the 700-mb chart is 3,010 meters, (9,880 feet) above mean sea level. This chart is used mostly to determine the vertical extent and structure of fronts and pressure systems, or to play the role of the 850-mb chart over elevated areas. It also plays the role of the 850-mb chart in moisture analysis in summer when moist tongues extend to greater heights than in winter, due to convective activity. Other uses of this chart are in forecasting (steering currents for certain shallow pressure systems are determined at this level) and in differential analysis. The contour interval is the same as the 850-mb chart.

Short waves are a predominant feature of this chart. They play a great role in the weather. For this reason the wave features of the 700-mb chart are carefully studied and tracked. Short waves have great influence on cloudiness, frontal intensity, precipitation areas, etc.

THE 500-MB CHART

The approximate height of the 500-mb level is 5,570 meters, (18,280 feet) above mean sea level. Primary features of this chart are the warm highs and cold lows with associated troughs and ridges. Long waves may also be identified at this level although most short waves have lost their identity.

This chart is the most widely used of all upper air charts for prognostic purposes. For various reasons, this chart comes closest to representing the mean state of the atmosphere at the time of observation. In addition to representing the wind structure at a common flight altitude (for piston-engine aircraft), it is also used extensively in forecasting the movement and development of fronts and pressure systems at low levels, sea level in particular.

Since this level approximately divides the atmosphere with respect to mass, it is often used in conjunction with the 1,000-mb chart to provide layer analysis in the lower half of the atmosphere. When no 200- or 300-mb chart is available, it gives a fair approximation to the horizontal position of the jetstream. Because the amount of data available decreases rapidly above this level, the 500-mb chart provides an important base upon which to construct high-level analyses. The contour interval is the same as for the 700-mb chart.

THE 300-MB CHART

The approximate height of the 300-mb level is about 9,160 meters (30,050 feet). The primary features of this chart are the permanent and semipermanent highs and lows, certain dynamic lows, long waves, the jetstream in winter, and the tropopause, especially the Arctic and midlatitude tropopauses in winter.

The primary uses of this chart are forecasting; determination of the characteristics of long waves; analysis and forecasting of the jetstream (with associated isotach analysis); analysis of the tropopause in winter; determination of the vorticity distribution; and in the case of tropical cyclones which do not show a closed circulation at this level, steering currents. Jet engines operate more efficiently at altitudes upward of 300 millibars; therefore, this chart, too, is an indispensable tool in planning jet operations. Contour interval is normally 120 meters; a 60-meter interval may be used in areas where a finer degree of delineation is required.

THE 200-MB CHART

The approximate height of the 200-mb surface is 11,790 meters, (38,660 feet). This chart is used principally as an adjunct to the 300-mb analysis. In summer, it plays the same role with respect to the jetstream as the 300-mb chart does in winter. In winter, its principal synoptic use is for estimation of advective temperature changes in the stratosphere. The 200-mb temperature analysis is particularly used in isotach analysis at 300 millibars. Operational use and contour interval is the same as for the 300-mb chart.

THE 150-, 100-, 50-, AND 25-MB CHARTS

These charts are normally prepared only by activities with large staffs and are used primarily for research purposes. The amount of data available at these levels is so scanty that analyses are likely to be greatly over-simplified and more than one logical analysis of the same data is easily possible. It can be anticipated that as the operational ceilings of jet aircraft increase, so will the practical uses of the 150- and 100-mb charts.

SPACE DIFFERENTIAL (THICKNESS CHARTS)

Space differential charts are commonly referred to as thickness charts, since they represent the difference in height between two constant pressure-surfaces. The most commonly used space differential charts are the 1,000- to 700-mb, the 1,000- to 500-mb, and the 500- to 300-mb charts.

They are usually constructed by graphically subtracting the heights on one analyzed constant pressure chart from those on another. The construction of these charts is covered in chapter 7 of this training manual.

Since vertical spacing between two pressure surfaces is directly related to the virtual temperature of the layer in question, the difference in height between two pressures is directly related to the mean virtual temperature of the layer. Consequently, liens of constant thickness on a space differential chart are also lines of constant mean virtual temperature for the layer. Space differential charts are an aid in the analysis and construction of prognostic charts.

The variation of wind with height is determined by the horizontal gradient of temperature. This difference in wind direction and speed from one level to another is referred to as the vertical wind shear or the thermal wind. The significance and determination of the thermal wind are covered in chapter 7.

CIRCULATION PATTERNS ON UPPER AIR CHARTS

The patterns on constant pressure charts take on much the same appearance as the patterns delineated by isobars on a surface chart. However, the pattern becomes smoother and more simple with an increase in elevation, since many of the pressure systems decrease in intensity with height. The 1,000-mb chart, for example, may show many closed centers, but by the time the 200-mb level is reached there are few if any closed centers. Instead, the contours present a wavelike pattern. These waves, like ocean waves have definite properties by which they may be identified.

LONG AND SHORT WAVES

Wave patterns are classified according to their wave lengths as long and short waves. Wave length is the distance from one trough to the next trough, or from one ridge to the next ridge, or from a point on one crest to the corresponding point on the next crest. Wave length is measured in degrees longitude. Wave amplitude is the north-south distance from point of inflection to peak or to lower extremity. Wave amplitude is measured in degrees latitude. Figure 4-14 illustrates the measurement of the length and amplitude of a long wave. Also note the appearance of a short wave trough on the long wave.

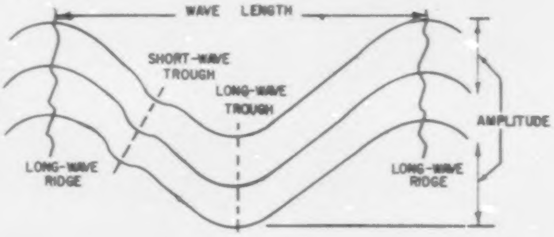

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Figure 4-14.—Illustration of a long and short wave and the measurement of length and amplitude of a long wave.

Long Waves

Long waves, sometimes referred to as major troughs, provide general guidance in prognosticating surface events, such as cyclogenetic areas and in determining steering currents for systems on surface synoptic charts. A long wave is a wave in the major belt of the westerlies which is characterized by large wave length and amplitude. Long waves vary in length between 50° and 120° of longitude. The amplitude of long waves is large and increases upward in the troposphere. Therefore, they are associated with cold troughs and warm ridges. Normally, the number of long waves around the hemisphere is 4 or 5, but there may be as many as 7 or as few as 3. Long waves move slowly with a normal movement at 40° N lat. on the order of about 2° long. per day during the spring to slightly less than 1° during the fall, but they can be stationary or even retrogress.

Long waves are persistent; that is, they do not appear or disappear rapidly. Formation generally takes place in the same geographical area when new waves are forming from short waves or from a changing synoptic situation. Long waves are easily identifiable from 5-day mean charts.

As related to synoptic events, long waves are associated with surface cyclone families. Over the eastern portion of the cyclone family may be found a major (warm) ridge; over the western portion, a cold trough associated with the polar outbreak. Therefore, major waves mirror the cyclone series in its entirety, while minor or short waves may be associated with individual surface cyclones. Long waves are best identifiable on the 300-mb chart, though during the colder season the 500-mb chart is frequently used for this purpose.

Figure 4-15 illustrates the relationship between long waves and other meteorological events.

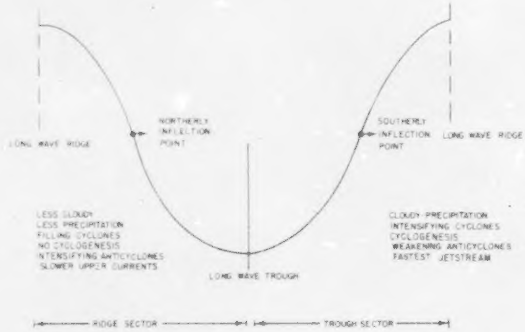

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Figure 4-15.—Long waves and related weather.

Short Waves

Superimposed on the long wave contours of a given upper air chart, say 500-mb, are numerous troughs and ridges of small dimensions called short wave troughs and short wave ridges. Short wave troughs are progressive waves of smaller amplitude and wave length than the long waves, moving in the same direction as the basic current in which they are embedded. They often disappear with height, and may not be detectable above the 500-mb level. Short waves are more numerous than long waves, with 10 or more active waves being present in the hemisphere most of the time. They are associated with warm troughs and cold ridges. Short waves move rapidly and are progressive. Their eastward motion is very near that of the 700-mb flow, and an average speed for these waves may be from 8° of longitude per day during the summer to 12° of longitude per day during the winter.

In comparison with long waves, short waves appear and disappear rapidly. Short waves appear to be of little consequence in the steering of surface pressure systems but do have a great effect on long waves. Short waves intensify as they approach long waves, and weaken as they leave the long wave behind. They flatten a long wave ridge when superimposed on one.

Because of the sparseness of upper air data in many areas, short waves may be easily overlooked or carelessly smoothed out of the analysis. In isolated areas, the principal clue to the passage of a short wave trough is a slight veering of the wind for a brief time. Too, the long wave trough intensifies when overtaken by a short wave trough and this intensification often results in surface cyclogenesis. One of the best synoptic clues to the existence and location of short waves in denser networks is the 12- or 24-hour coincidence of small closed height fall centers with short wave troughs and the 12- or 24-hour small closed rise centers with short wave ridges.

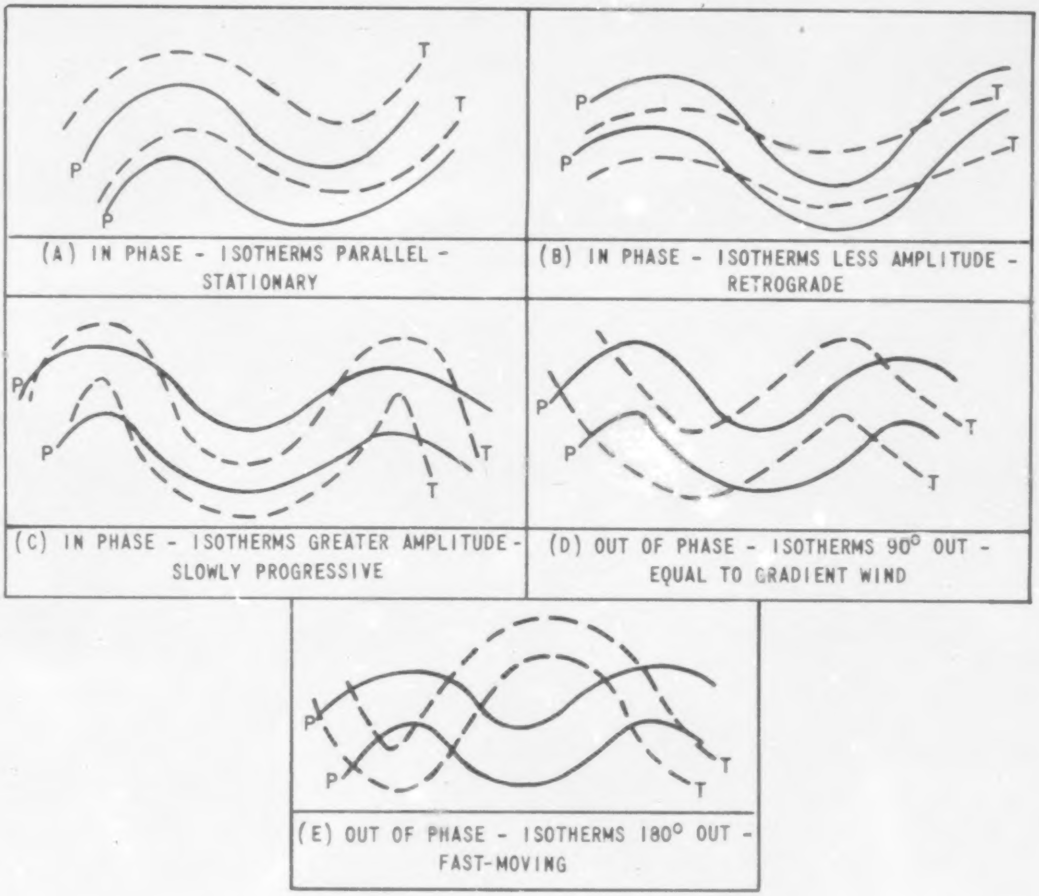
Isotherm-Contour Relationship

Certain generalized patterns of isotherms in relation to contours are of significance in that they are indicative of present conditions, hence, future conditions also, especially in relation to movement of troughs and ridges.

Isotherms on constant pressure charts are considered in phase or out of phase with contours. When isotherms are in phase, they form a pattern in relation to the contours that is exactly coincident with the troughs and ridges. They are also deemed in phase when the isotherms are of greater or lesser amplitude than the contours. Isotherms are out of phase when the trough and ridge lines of the isotherms are not coincident with the contour trough and ridge lines. (See fig. 4-16.)

The rules for the movement of waves determined by the isotherm contour patterns are as follows:

1. If the isotherms are in phase and parallel to the contours, the wave is stationary.
2. If the isotherms are in phase but of less amplitude then the contours, the wave will have retrograde motion.
3. If the isotherms are in phase but of greater amplitude than contours, the wave is slowly progressive.



(A) IN PHASE - ISOTHERMS PARALLEL - STATIONARY

(B) IN PHASE - ISOTHERMS LESS AMPLITUDE - RETROGRADE

(C) IN PHASE - ISOTHERMS GREATER AMPLITUDE - SLOWLY PROGRESSIVE

(D) OUT OF PHASE - ISOTHERMS 90° OUT - EQUAL TO GRADIENT WIND

(E) OUT OF PHASE - ISOTHERMS 180° OUT - FAST-MOVING

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Figure 4-16.—Isotherm-contour patterns.

4. If the isotherms are 90° out of phase, the wave moves with the speed of the gradient wind.

5. If the isotherms are 180° out of phase, the wave is fast moving.

UPPER HIGHS AND LOWS

The most common pattern which appears on upper air charts consists of alternate troughs and ridges, with occasional closed lows in the troughs and closed highs in the ridges, forming a series of waves which girdle each hemisphere. The lows are normally on the poleward side of the strongest westerlies, and the highs are on the equatorward side of the strongest westerlies. These lows and troughs generally are, but not necessarily, associated with cyclones at lower levels (sea level in particular).

If the situation is reversed, that is a low appears on the equatorward side of the jet and highs are poleward, they are said to be in an abnormal position. Highs and lows in this abnormal position are sometimes referred to as CUTOFF CENTERS. They are most common in the spring and least common in the fall. In seasons when they are most frequent, they repeatedly occur in the same geographical location. Cutoff lows appear most frequently over the southwestern United States and along the northwestern coast of Africa. Cutoff highs are most frequent in northeastern Siberia, Alaska, and Greenland.

The formation of these cutoff centers is discussed later in this chapter. The periods in which these occur is believed to be part of the index cycle taking some 4 to 6 weeks to accomplish. When highs and lows are in this position, we appear to have our topsy-turvy weather. For instance, the temperature may be warmer in Alaska than in Florida.

SPECIAL CIRCULATION FEATURES

During the discussion of secondary circulation you noted that centers of action migrate and change in intensity. These changes in location, orientation, and intensity of the centers of action influence the movement of migratory systems. Since special circulations play a large role in influencing the secondary circulations, it is important that you learn the significant features of these circulations and how they can influence the daily, and sometimes the seasonal, weather.

ZONAL AND MERIDIONAL FLOW

Variation in the speed and direction of the west wind is one of the most significant aspects of the atmospheric circulation. Probably, the first, and certainly the simplest, of all methods of classifying the flow patterns is according to whether west-east or north-south components of flow dominate. The former case is characterized as zonal flow; the latter, as meridional flow. Figure 4-17 shows typical examples of zonal and meridional flow on the 500-mb chart.

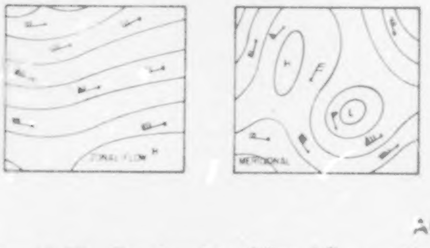

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Figure 4-17.—Zonal and meridional flow patterns.

Under zonal conditions, the long waves have smaller amplitudes and longer wave lengths than they have under meridional flow. Another way of characterizing the long wave pattern is by wave number. Zonal flow has a small wave number; meridional flow a large wave number. Surface systems are comparatively shallow (in vertical extent) and move rapidly west to east in zonal patterns, while in meridional patterns they are deeper in vertical extent and move poleward or equatorward, or remain quasi-stationary.

ZONAL INDEX

The zonal index is a measure of the strength of the midlatitude westerlies, expressed as the horizontal pressure difference between 35° and 55° latitude, at the surface or aloft. The upper level most frequently used (and the most convenient to use) is the 700-mb level.

A quantitative expression of the index is associated with certain weather phenomena. The zonal index does not, however, account for dynamic processes in the upper air such as convergence and divergence; consequently the zonal index is now generally used only as an indicator and is seldom computed, but merely estimated.

The zonal index may be computed at the surface by averaging the reported and estimated sea level pressures at latitudes 35° and 55° at 5° longitude intervals over a span of at least 90° but preferably 120°, longitude. The mean sea level pressure at 55° latitude is then subtracted from the mean sea level pressure at 35° latitude. If the sea level pressure difference is at least +8 millibars, high zonal index prevails; and if the difference is +3 millibars or less, low zonal index prevails.

When computing the zonal index at 700-mb level, the actual contour values are used, and the final difference is then converted to millibars by letting 60 meters equal 7 1/2 millibars.

High Zonal Index

When the pressure difference between 35° and 55° latitude is more than 8 millibars, high zonal index prevails. High zonal index is associated with long waves in the upper air, having small amplitudes, warm troughs, and cold ridges.

With high zonal index the Icelandic and Aleutian lows are in normal position, or slightly north of their respective normal positions and well developed. The axis of these lows will be east-west, and the associated troughs will have an east-west orientation. The Atlantic and Pacific subtropical highs are north of their respective normal positons and their orientation is strongly east-west. The Great Basin High is present, as is the Siberian High, but it has little westward extension. Highs are absent in the high latitudes. Frontal zones move from west to east, are at higher latitudes, move rapidly, and have a prominent east-west orientation. There are few polar outbreaks, and the moderate highs in the low latitudes move rapidly. The midlatitude temperatures are moderate, and the weather is generally fair. The polar regions become colder and at 60° N lat. the weather is stormy.

Low Zonal Index

Low zonal index prevails when the sea level pressure difference between latitudes 35° and 55° is +3 millibars or less. An extreme low zonal index is at hand when the pressure at 55° latitude is higher than that at 35° latitude. On the 700-mb chart short waves are a predominant feature. The short waves have relatively large amplitudes and are associated with warm ridges and cold troughs. Cutoff centers are common. The Icelandic and Aleutian lows have split into two weak cells each, and they have a strong north-south orientation. The Atlantic and Pacific subtropical highs are weak, and in a more southerly position than normal with a north-south orientation. Each of the highs has split into two cells. The polar region highs are strongly developed, and joined in the Arctic regions. There are more fronts during low zonal index; there is sharp contrast in them; their orientation is north-south. Storms and cyclones with heavy precipitation are frequent in the low latitudes. Temperatures occur in extremes in the mid and low latitudes. The high latitudes are storm-free with mild temperatures.

Changing Zonal Index

The zonal index does not change from high to low overnight; however, the degree of high or low zonal index is not immediately a static value. The weather situation changes irregularly from high to low zonal index and vice versa. Changing zonal index is therefore an important aspect to consider in estimating the current trends. A changing zonal index may be a falling or rising zonal index.

FALLING ZONAL INDEX.–The Icelandic and Aleutian lows move southward and begin to split. The Atlantic and Pacific subtropical highs follow course. Migratory systems slow down, especially in the higher latitudes, resulting in a gradual shift in frontal orientation to north-south. Polar highs develop and begin to break out.

RISING ZONAL INDEX.–The eastern cells of the semipermanent lows and highs begin to fill and weaken, respectively, and move eastward, finally disappearing completely. The western cells of the centers of action move northeastward into their normal positions. Migratory systems begin to intensify and speed up, especially in the higher latitudes, resulting in a more east-west orientation of the associated frontal systems. The polar high stagnates over the Great Basin; cutoff systems begin to fill.

The seasonal change in the zonal index is of great magnitude. Summer changes are small and winter changes are extreme. A particular index may remain nearly stagnant for several weeks, especially in the winter, or manifest itself only for a few days.

BLOCKS

Blocking is defined as the development of a warm ridge or cutoff high aloft at high latitudes which becomes associated with a cold high at the surface. Blocks form most frequently in the northeastern portion of the Atlantic and the Pacific. They move very slowly and tend to move westward during intensification and eastward during dissipation. Atlantic blocks frequently move as far westward as the Maritime Provinces of Canada. The blocking high is the situation when a warm core upper high has become situated over a cold core surface high, splitting the westerlies.

The cutoff high is a particular example of the general class of highs called blocking highs. A normal accompaniment sufficient for anticyclogenesis is the existence of high-speed wind approaching anticyclonically curved and weaker contour gradients downstream. The highest pressure to the right of the high-speed (jet) winds tends to be propagated downstream to fill and cause to move east any trough in its path downstream.

Three general types of blocks are said to exist. One is called the OMEGA block due to its resemblance to the Greek letter OMEGA. Figure 4-18 shows the three types of blocks as they occur in the atmosphere.

Identifying Characteristics

For identification purposes, blocks have the following characteristics: The basic westerly flow splits into two branches, and each branch transports an appreciable amount of mass; about half the contours go north and the other half south. Each branch must extend over at least 45° long. before they recombine; a sharp transition from zonal to meridional flow must exist in the region of the split; upstream the flow is zonal and downstream meridional.

There are other characteristics for recognition and detection. The high is centered in middle or high latitudes (poleward of 40° lat.). There is a strong jet or branch of the jetstream going around the poleward side of the high. They tend to occur persistently in the same season or not at all. They occur most frequently in late winter and early spring.

Isotherm-Contour Relationship

In a typical blocking pattern, a core of warm air is present in the vicinity of the cutoff high; if a low is also present, a cold air core is present in the vicinity of the cutoff low.

Formation

The genesis of a typical blocking situation is normally preceded by intense cyclogenesis to

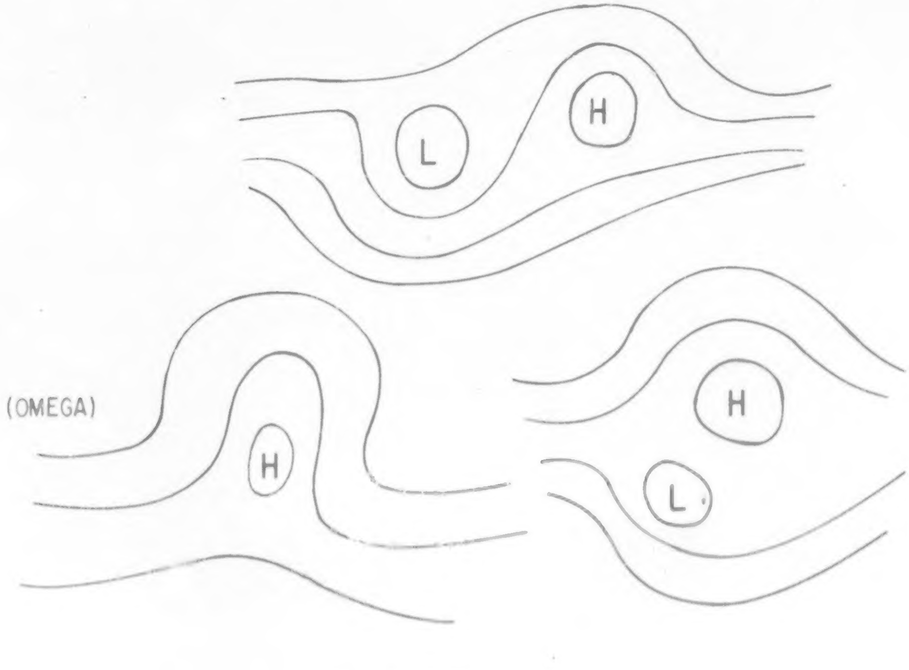

Figure 4-18.—Types of blocks.

the west and the development of a cold continental anticyclone eastward of the blocking area. This brings about induction of warm air into the ridge, causing the warm ridge to cut off and split the westerlies. Several days are normally required to accomplish this.

Blocks should be looked for in late winter or early spring in certain preferred locations; that is over the eastern portions of oceans in high and middle latitudes. During intensification, blocks have a tendency to move slowly westward (retrograde). This movement is very slow.

Dissipation

The reestablishment of the strong zonal flow by whatever process it is induced causes the dissipation of blocks. During such dissipation processes, blocks tend to move eastward at a very slow rate.

CUTOFF LOWS

Waves in the contour patterns on upper air charts can be either stable or unstable, the same as waves on a front, depending on whether their amplitudes increase with time. Figure 4-19 shows the typical stages in the development of a CUTOFF LOW in the trough of an unstable wave.

Stages 3 and 4 correspond to occlusion in the wave cyclone and are often associated with surface occlusions. Cutoff lows are always dynamic. Cutoff highs are defined in a similar fashion and are also dynamic. Because of the obvious analogy, some meteorologists have suggested that a surface high associated with a cutoff high aloft be called an occluded anticyclone.

Cutoff lows occur most frequently along the southwestern coastal area of the United States and the northwestern coastal areas of Africa. They may form just south, but mostly southwest, of a blocking high. A very sharp ridge is usually to the west of them, oriented southwest to northeast at 500 millibars. The cutoff low forms a pool of cold air. It is a completely isolated cyclone of cold air. The conditions, as illustrated in figure 4-14, leading to the formation of a cutoff low are high-speed winds and height falls downstream approaching weak cyclonically curved contours of a low in the midlatitudes. The direction of these high-speed winds is such that in deepening, the low is also displaced farther south or remains quasi-stationary (at least in the formative stages).

THE JETSTREAM

A jetstream is a band or belt of strong winds of 50 knots or more with a strong westerly component which meanders vertically and horizontally around the hemisphere in wavelike patterns. It is not found at the same latitude or elevation around the earth, but has a sinusoidal pattern (a wavelike trajectory). At times it is a continuous band, but more often it is broken up into several discontinuous segments, or it is split at several points. Jetstreams are characteristic of

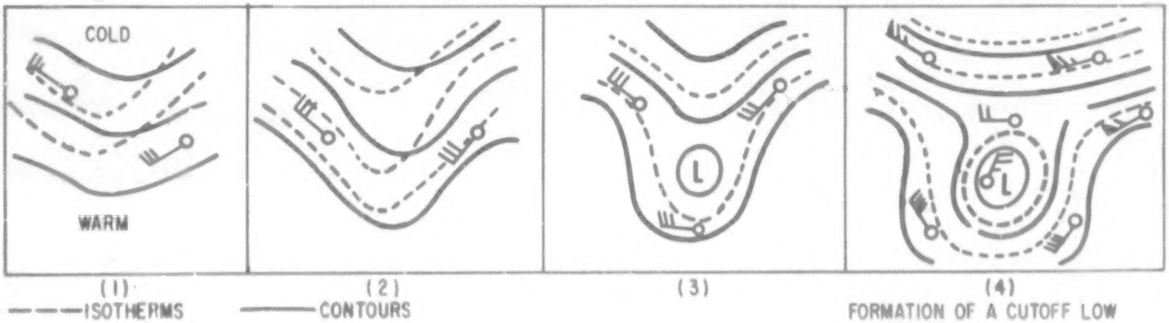

Figure 4-19.—Development of a cutoff low.

both the Northern Hemisphere and the Southern Hemisphere, however, much more is known about the ones in the Northern Hemisphere, since there is more data available about them. The general characteristics mentioned here are based on what is known of jetstreams in the Northern Hemisphere.

An observation of high wind speed does not by itself warrant use of the term "jetstream." It is necessary for large vertical and horizontal shears to exist which limit vertical and lateral extent of the strong winds, in order for a "core to exist." Moreover, the word "stream" implies that the core must possess considerable length.

For the sake of definition, the World Meteorological Organization in 1958 proposed the following definition: "Normally a jetstream is thousands of kilometers in length, hundreds of kilometers in width, and also some kilometers in depth. The vertical shear of wind is of the order of 5 to 10 m/sec per km and the lateral shear is of the order of 5 m/sec per 100 km. An arbitrary lower limit of 30 m/sec is assigned to the speed of the wind along the axis of the jetstream." The rate of change of wind direction and speed with either altitude or on a horizontal plane is termed shear.

If one visualizes a current with dimensions as given by the WMO, he finds that the jetstream can be likened to a thin, narrow ribbon. A system of such shape cannot be portrayed on true scale on vertical cross sections where a description of vertical and horizontal gradients is desired. On such sections the vertical distance is always exaggerated, usually about 100:1. This scale distortion must be kept in mind when cross sections are examined.

One jetstream rarely occurs alone in the atmosphere; multiple jets are more the general rule. They have been found to have a pronounced association with the subtropical high and frontal systems that occur between selected air masses; the Arctic or "Polar Night" jetstream is not associated with any of these phenomena, but is situated at high altitudes in the Arctic at an altitude where the stability of the atmosphere increases discontinuously upward.

In general, jetstreams are labeled as the polar front jet and the subtropical jet; a later discovery is the Arctic or "Polar Night" jet. However, for the purpose of identification and forecasting a still finer delineation is necessary. For instance, the polar front jetstream may be either a single or multiple system, depending on certain meteorological factors. Listed below are the principal divisions of these jets:

1. Polar Front Jetstream (Single System). In both cases of the polar front jetstream it is associated with the principal frontal zones and cyclones of middle and subpolar latitudes. The polar front jetstream as a single system is a characteristic occurrence during a high index condition.

2. Polar Front Jetstream (Multiple System). During low index conditions, the air mass discontinuity associated with the polar front is modified, and as a result a branch of the polar front jet system is established north and west of the migratory high in addition to the old polar front branch that remains south of the high. This is especially true over the Pacific through the action of a migratory polar high from Japan which stagnates in the Pacific. The northern branch of this jet system has been labeled the interpolar front jet; the other, the old polar front jet. Once the formation of the interpolar front is initiated, the old polar front jetstream is apt to disappear rather rapidly and the interpolar front jet becomes the dominant one.

3. Subtropical Jetstream. This jetstream is found solely in the lower latitudes over subtropical high cells and marks the poleward limit of the trade wind cell of the general circulation. Generally, the transfer of angular momentum is assigned the leading role in the formation and maintenance of this jetstream. Thermal contrasts are in evidence on the left of the stream (looking downstream), though they are not as intense as those contrasts between air masses.

4. Arctic or "Polar Night" Jetstream. This jetstream is situated high in the stratosphere in and around the Arctic or Antarctic Circles.

Although there may be other currents resembling jetstreams at still higher levels, our knowledge about them is still too scant to warrant treatment of them here. Although the three types of currents occur in widely different geographic locations, their basic structure is similar. The basic structure of the jetstream is discussed in this chapter and differences between the polar jet and the other two pointed out where appropriate.

Synoptic Structure of the Jetstream

The model of the jetstream in the upper air in common usage shows a strong concentration of high wind speed lines (isotachs) both vertically and horizontally in the upper air that are organized into fairly narrow ribbons embedded in the relatively slow speed wind fields of the surrounding atmosphere. The concentration is located just to the right of the baroclinic zones (looking downstream) between relatively cold and warm air. A complete representation of the wind structure should take into account its variations in three dimensions—vertically (depth), latitudinally (width), and longitudinally (length). Since the wind field in a three-dimensional picture is hard to portray, a two-dimensional picture is used. They are the vertical and horizontal fields. (See fig. 4-20.)

Thermal Field Around the Jetstream

A jetstream (other than the subtropical jetstream) is associated with the thermal discontinuity between air masses. The subtropical jetstream does not depend upon air mass contrast but rather on momentum transfer.

The magnitude of the thermal discontinuity between the air masses depends upon the initial properties of the individual air masses and the degree of subsequent modification of these two properties.

One of the most valuable clues in locating the jetstream stems from the relationship of the jetstream to the thermal discontinuity (front) between two air masses. The core of the jetstream is located directly above, or nearly so, the thermal concentration on the 500-mb surface. Since recent studies have revealed that troughs and ridges slope very little from 500 to 300 millibars and are almost vertical from 300 to 50 millibars, a high degree of correlation is evident in locating the jetstream on charts above the 500-mb level.

Frontal intersections aloft appear on the horizontal planes as areas of highly concentrated temperature lines or zones of maximum concentration of temperatures. This concentration of isotherms may reach a magnitude as great as 10°C in 45 miles, but usually ranges between 10°C in 90 miles and 10°C in 150 miles. Under certain conditions this zone of concentration may be traced continuously around the hemisphere, but not often. The degree of concentration will vary along the front, being weaker, in general, through ridges and stronger in troughs.

Since a concentration of temperature represents the frontal intersection, more than one zone of concentration would exist if more than one front extends to any one plane. This is particularly evident at low levels, 500 millibars and below. Above 500 millibars only the major frontal systems are likely to appear.

A jetstream is usually associated with each thermal concentration in excess of 10° C in 200 miles, expecially if the thermal contrast is maintained through a deep layer. This thermal contrast comprising the frontal zone is usually evident at all levels up to and including the 400-mb level, and, on occasions as high as the 300-mb level. A positive thermal gradient (warm air to the south and cold air to the north) is present in these levels. Between 300 and 200 millibars their thermal contrast is greatly diminished or disappears. Above the 200-mb level their thermal contrast is reestablished, reflecting the appearance of warm stratospheric air to the left of the jetstream (looking downstream) and cold tropospheric air to the right. Thus in these high levels, the direction of the thermal gradient is now reversed (warm air to the north and cold air to the south). The level at which the reversal of temperature gradient occurs is termed the Z-SURFACE. The foregoing temperature distribution is necessary if the wind is to increase to a maximum value and then decrease. (See fig. 4-20.)

Thus, it may be said that the jet core will lie between 200 and 300 millibars directly above the strongest meridional temperature gradient at 500 millibars. A strong meridional temperature would be indicated on the 500-mb chart by a packing of the isotherms.

Location on Upper Air Charts

From these observations we can evolve the following rules: With the exception of the subtropical jetstream, expect jetstream cores (especially midlatitude jets) to be found directly above, or nearly so, a zone of maximum concentration of temperatures on the 500-mb

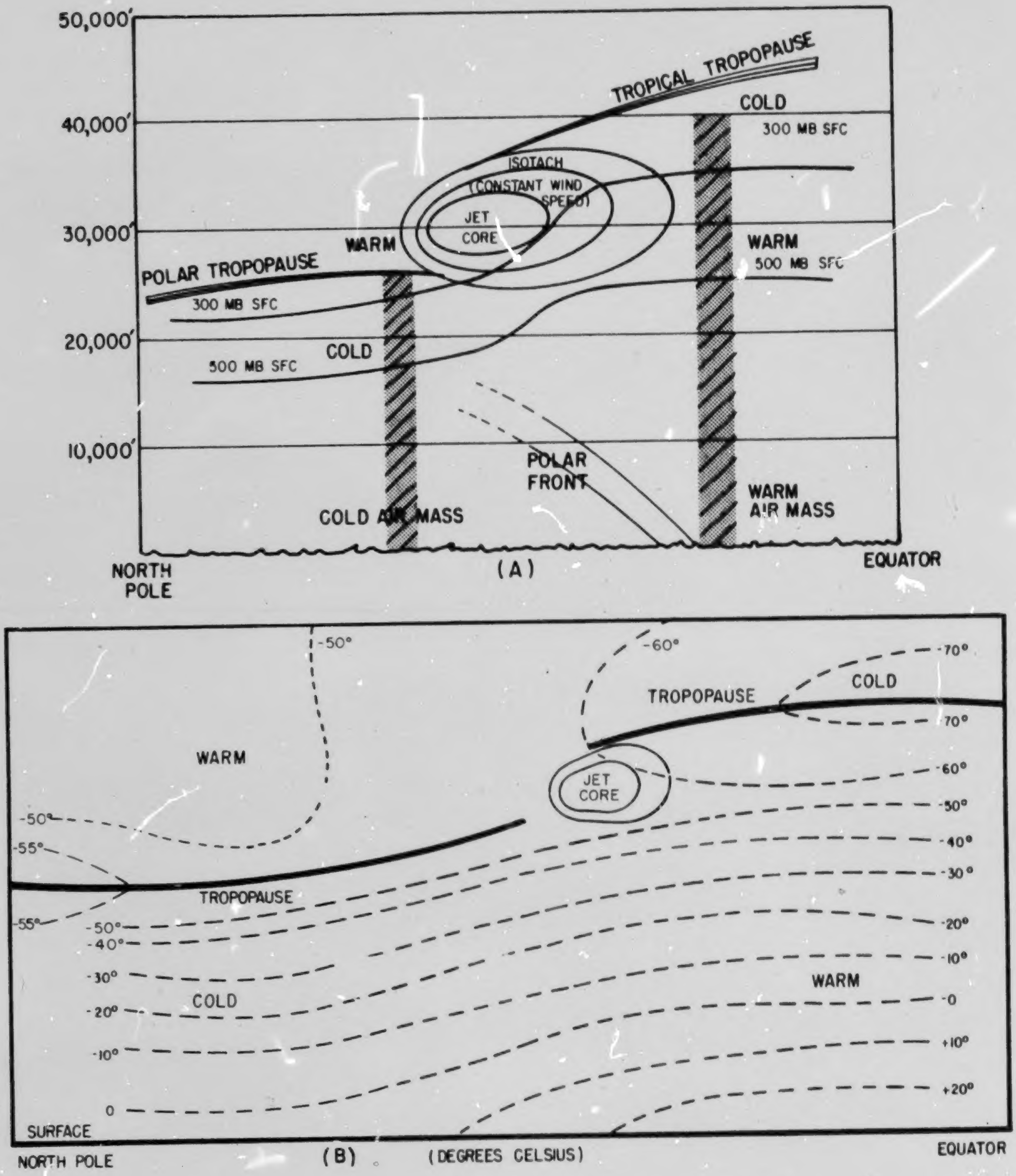
50,000'
40,000'
30,000'
20,000'
10,000'
TROPICAL TROPOPAUSE
COLD
300 MB SFC
ISOTACH
(CONSTANT WIND SPEED)
JET CORE
WARM
POLAR TROPOPAUSE
WARM
500 MB SFC
300 MB SFC
COLD
500 MB SFC
POLAR FRONT
WARM AIR MASS
COLD AIR MASS
NORTH POLE
(A)
EQUATOR
-50°
-60°
-70°
TROPOPAUSE
COLD
70°
WARM
JET CORE
-60°
-50°
-55°
-50°
-40°
-30°
TROPOPAUSE
-55°
-50°
-40°
-20°
-30°
-10°
WARM
-20°
COLD
-0
-10°
+10°
+20°
0
SURFACE
NORTH POLE
(B)
(DEGREES CELSIUS)
EQUATOR

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Figure 4-20.—Vertical cross section of a model of the jetstream. (A) Wind and fronts; (B) temperature distribution.

surface. O'Connor states that the jetstream is frequently well defined at this level and is usually found in the −17° C to −20° C isotherm ribbon with a most frequent concentration along the −17° C isotherm. The contour channels most frequently associated with the jetstream are between 10,200 (5,548 meters) and 18,600 feet (5,670 meters), with a concentration along the 18,400-foot (5,608 meters) contour. Project AROWA found in a research study that the jet maximums occur 82 percent of the time between 29,400-foot (8,960 meters) and 30,300-foot (9,236 meters) contours. This should be primarily a wintertime rule, since the height would vary with area and season and specified climatic conditions. Also, from the foregoing we can expect the width of the core to be approximately equal to the width of the zone of maximum concentration of temperature on the 500-mb surface.

Relation of the Jet to the Polar Front

Since the jetstream is located above the polar front at 500 millibars, we could expect the core of high speeds to be almost directly overhead at a station where the frontal intersection is found close to the 500-mb level at that station. If the frontal intersection is found below 500 millibars over a station, the core of high speeds will be north of the station. When the frontal intersection is found above 500 millibars above the station, the core of high speeds are found south of that station.

While a thermal concentration is usually quite apparent at levels up to and including the 350-mb level, there is one important exception, the Arctic front. Due to the shallowness of the air masses involved, this particular frontal discontinuity is not often apparent on the 500-mb surface.

Distribution of the Wind Field through the Jetstream

The wind field comprising a jetstream is so distributed that a cross section taken normal to the windflow reveals a system of annular speed lines (isotachs). (See fig. 4-20.) In a well-organized jetstream there is a nonuniform spacing of these isotachs around the core such that a concentration is greatest through the frontal area and above the core.

With the exception of the shear through the frontal zone, the shear above the jet core (in close proximity to the tropopause) is much greater than the shear below the jet core.

VERTICAL SHEAR.—Within the boundaries of the jetstream are found the strongest vertical wind shears observed in the belt of the westerlies. In vertical cross sections, the isotachs north and south of the core are very close to being vertical, indicating little variation of wind speed with height in these areas. In general, the highest values of vertical wind shear are encountered above the 500-mb level. In summer and sometimes in winter, the strongest vertical wind shears may be located above 300 millibars and become negligible near 500 millibars. Thus at 500 millibars, winds at this level may not give an indication of winds at high elevations.

HORIZONTAL SHEAR.—Vertical cross sections of the jetstream also indicate horizontal wind shear. There is a rapid decrease on either side of the jet core. In extreme cases, as much as 100 knots in 100 miles on the north side and 100 knots in 300 miles on the south side of the core will occur. The greatest cyclonic and anticyclonic shears are located at the level of maximum winds. Cyclonic shear is found poleward of the jet axis; anticyclonic shear, equatorward.

The jetstream follows the configuration contours on upper level charts. In view of the variation of the level of the jetstream with latitudes (low at high and high at low latitudes) and its known tendency to shift to higher elevations in the warmer season of the year, the 300-mb chart is considered suitable for locating the jet at high latitudes in winter and the 200-mb chart for middle latitudes. This is due to the fact the polar jetstream core is usually located between 300- and 200-mb levels. A higher chart is used for locating the subtropical jetstream as it is usually located somewhere between the 200- and 100-mb level, and is often near the 150-mb level.

Since the jetstream varies with elevation along the line of flow, it does not always appear as a continuous flow on a constant pressure chart.

General Remarks About the Jetstream

Jetstreams undergo a distinct life cycle consisting of a period of organization and one of disorganization. Their horizontal structure, as it appears on upper level charts, depends on the phase of this life cycle. When the jet is well organized, the longitudinal axis follows the pattern of long waves, and its amplitude is comparable to that of the 300-mb contours. A lack of uniformity of wind speed along the axis is very evident: there are intervals where the winds are comparatively weak. This creates a rather sharp wind wind gradient along the longitudinal axis, for the difference between wind speeds at maximum and minimum points may amount to over 100 knots.

During disorganized periods, the well-defined concentration of speeds is missing, and these west winds either break up in circular flow, or show fairly uniform speeds throughout the mid latitudes. It is during this period that numerous splits or JET FINGERS appear. These fingers are all comparatively weak and sometimes separated by as little as 5° latitude.

Centers of high and low wind speed alternate along the axis of a well-developed jetstream. The maximum speeds in these centers may reach well over 250 knots. The distance between successive maximums is not constant, but varies from about 10° to 25° longitude. Variations of height of these maximums also exist along the axis. These maximums are associated with short wave troughs in the westerlies and move with a speed proportional to the 700-mb flow.

Multiple Jets

The concentration of west wind in the middle and upper troposphere does not occur along a single axis. On almost any day maps reveal the presence of at least two bands of speed concentration. Occasionally three or four distinct bands are in evidence for short periods. The individual jet bands tend to lie parallel to the upper contours, but appreciable departures from this alinement are often observed, especially east of the trough. The different jets are not of equal intensity. The most intense is usually the midlatitude jet and is associated with the zone of greatest temperature gradient in the low troposphere and mid-troposhpere. These jets axes then, have a tendency to attain their maximum strength in midlatitudes. Individual jet axes often show a net equatorward shift of about one-half degree of latitude per day. New axes form in high latitudes as the older ones move out. The equatorward migration is usually accompanied by a rise of the core to higher elevations. Although the separate jet bands usually retain their identity around the hemisphere, they occasionally combine into one or more regions. Conversely, a single well-defined jet is often observed to split into two or more branches. In this case, the northern branch often shows a temporary northward trend. This split is associated with blocks.

Relation of Jetstream to the Tropopause

Since the level of reversal of the temperature field often coincides with the tropopause, attempts have been made to locate the level of strongest winds from tropopause analysis. This has proved a difficult undertaking. The tropopause often lies above the level of maximum wind and is therefore not a perfect indicator for the maximum wind core. Besides, tropopause analysis is very complicated because a single and simple tropopause is rarely observed in middle latitudes. More often, the tropopause is ill defined or there is a multiple structure, and many assumptions and definitions of an arbitrary character are necessary in drawing tropopause analyses.

One method of associating the jetstream with the tropopause has received rather wide acceptance, but it is not completely representative of all cases for the reasons stated above. The jetstream is assumed to be associated with a break in the tropopause. North of the jet the tropopause is low, and south of the jet the tropopause is high. If the tropopause is continuous through the jet, it has a very sleep slope in the region of maximum wind. Oftentimes, the two tropopauses overlap with the jet located between them. The midlatitude or polar front jet is associated with the break between the subarctic and midlatitude tropopauses (30° to 50° N lat.). A second jet, the subtropical jet, is frequently associated with a break between the midlatitude and subtropical tropopauses (25° to 30° N lat.).

Associated Clouds and Weather

Statistical and observational studies of cloud patterns associated with the jetstream reveal that most of the time there are no clouds at or above the level of the jetstream core. In these studies, however, cirrus generally reached to within several thousand feet of the tropopause on the equatorward side of the axis and no clouds were observed above the tropopause there.

In one study, clouds were most frequent in the upper troposphere with the center of cloudiness 10,000 to 15,000 feet below the core and from 4° to 5° poleward of the core. Also, it was revealed that another center of cloudiness existed at about the same distance equatorward at about 5,000 to 10,000 feet below the level of strongest wind. On many occasions the crew members of research aircraft engaged in a study of cloud patterns associated with the jetstream reported a sharp discontinuity in the cirrus near the core with cloudless skies to the immediate north. The observers also reported that in cases where the cirrus extended to the north of the core there is often a narrow break in the cirrus at the core itself. The thickness of the cirrus averaged about 500 feet, but individual layers ranged from a hundred to several thousand feet in depth.

There are generally recognized to be four cloud patterns associated with the jetstream. They are as follows:

1. Lines of cirrus in bands (H4, 5, and 6).
2. Patches of cirrocumulus (H9) or altocumulus castellanus (M8).
3. Lenticular clouds in waves (M4 or M7).
4. Waves of altocumulus (M3 or M5).

At least three of the four patterns must be present for a jetstream to be justified, and they must present a pattern. These cloud patterns often extend from the horizon to the horizon, and have waves at right angles to the airflow. From the ground, they are observed to move at high speeds (except the lenticular type), often resulting in rapid local changes in cloud cover over a short period of time. They show streaks in the direction of the wind and occur in coherent patterns. The use of the cloud method in identifying or detecting the presence of the jetstream is restricted to periods when lower clouds do not obscure the sky and when enough moisture is present in the jetstream to permit cloud formation. These cloud patterns can be attributed to the convergence occurring downstream from the jet maximums and to the lifting created by the undulatory motion in the areas of wind shear along the jet axis.

In view of the wide variety of jetstream structures which occur, precipitation occasionally will be encountered in almost any position with respect to the high speed center.

But as with cloud formations, the occurrence of precipitation associated with the jetstream is controlled primarily by the distribution of wind shear and curvature along the jetstream.

Several studies have been made relating precipitation to the jetstream. In all, there is general agreement that the highest incidence of precipitation almost straddles the jet axis, with a slight bias toward the poleward side. However, in one of these studies it was found that precipitation occurred only 50 percent of the time near the axis, and that large portions of the core were without precipitation.

Climatology of the Jetstream

From day to day the jetstream can move in a direction at right angles to its axis, with a speed of approximately 10 knots. This motion can be either toward the polar air or toward the tropical air.

The mean position of the jetstream shifts south in the winter and north in the summer, with the seasonal migration of the polar front. The average latitude of the jetstream in the Northern Hemisphere in summer is 42° N lat; in winter it is 25° N lat. Additionally, there is a climatological variation in longitude, especially in winter. High wind speed concentrations are found near the east coasts of continents and lower wind speed concentrations are found on the west coasts of continents. The rule is that speed concentrations increase as the westerlies pass over the continental anticyclonic circulations and weaken as they pass over the oceans.

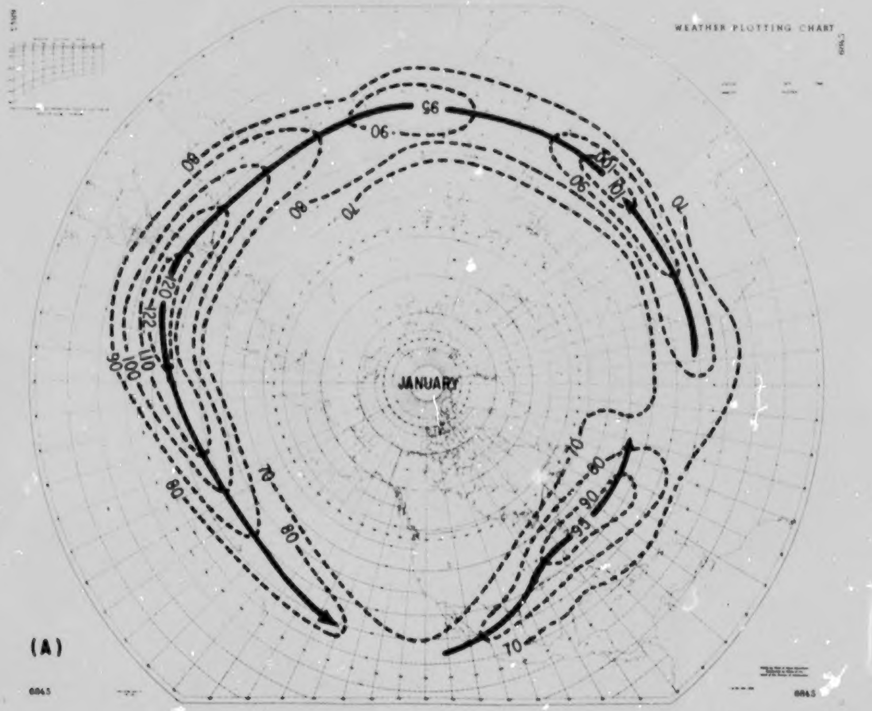

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Figure 4-21.—Mean seasonal jetstream distribution. (A) January.

See figure 4-21(A) and (B) for the mean seasonal positions of the jetstream during January and July.

Latitudinal fluctuations of the jetstream are greatest on the west coast of continents.

Relation of the Jetstream to Fronts and Cyclones

Jetstreams pass over thousands of miles of subtropical desert areas where there is little or no cloudiness and where the surface circulation is anticyclonic. They also lie above the strongest cyclones in the westerlies with extensive sheets of clouds and precipitation extending along the axis. Thus, the mere existence of a jetstream in any area, even a very strong one, does not of itself imply bad weather there. The converse, however, holds to a much better degree of approximation, especially in winter; cyclones and extensive bad weather areas tend to be connected with jetstreams. This is especially true for weather associated with a warm front in the formative and mature stages of a cyclone.

We have stated previously that the midlatitude jet usually lies directly above the polar front at 500 millibars. Using an average frontal slope and an average height of the 500-mb surface (18,280 ft), you can place the jetstream some 300 miles behind the surface cold front and around 600 miles in advance of a warm or stationary front. However, in estimating the location of a front at 500 millibars in mountainous areas, the height of terrain must be considered. For example, if the height of the land is 10,000 feet, the jetstream should be placed only about 300 nautical miles in advance of the surface warm front.

A strong surface front does not guarantee a jetstream aloft, as a strong temperature contrast

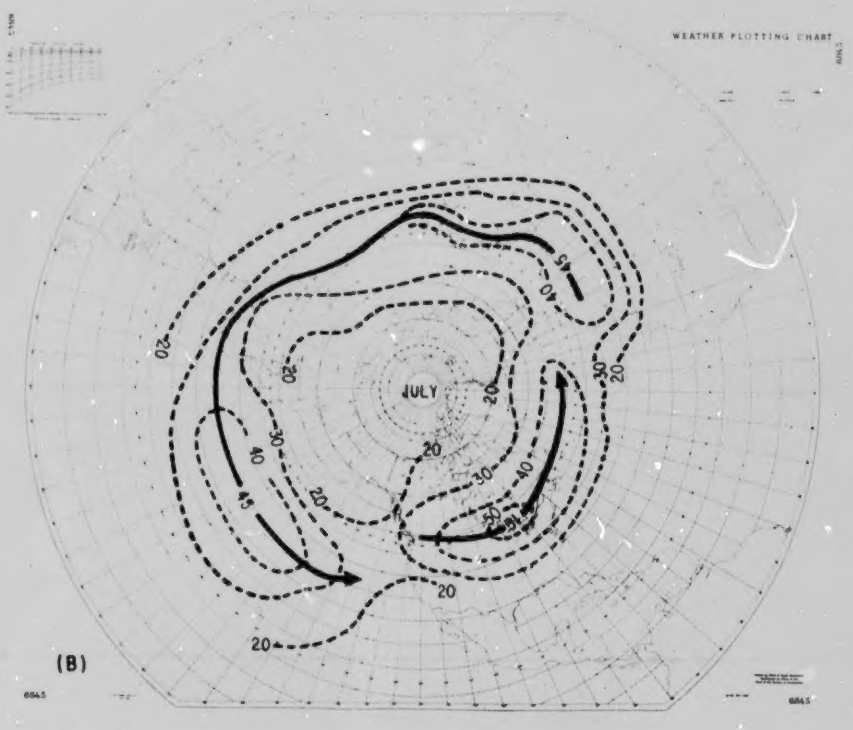

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Figure 4-21.—Mean seasonal jetstream distribution—Continued. (B) July.

throughout a deep layer is required. In some cases, a wedge of cold air behind a front will be overrun by a warm southwesterly current, producing extensive cloudiness and continuous precipitation. This should be taken as definite evidence that the cold wedge is shallow. The jetstream behind such a front will be displaced well to the north, usually about 600 miles.

If there is enough moisture to produce typical cloud patterns with fronts, the location of the front at 500 millibars may be estimated as being near the transition from high to middle clouds. The jetstream will usually lie directly above this transition zone.

The relationship of the jetstream to surface lows and fronts is given in the following section:

1. The jetstream will be perpendicular to an occlusion and to cold fronts oriented north-south with no associated warm front.
2. The jetstream will remain north of an unoccluded wave cyclone.
3. The jetstream will be south (near the point of occlusion) of the low associated with an occluded front.
4. In a series of lows of a cyclone family each low will be associated with a jetstream maximum. (See fig. 4-22.)

The two preferred positions of a low center with respect to a moving jetstream maximum are given in figure 4-22. As the wave cyclone in the right rear quadrant (looking downstream) deepens, the associated tightening of the thermal wind (and contour) gradient produces a new wind maximum. Further deepening and occlusion displace this maximum south of the low placing the low in the left front quadrant of the jet maximum.

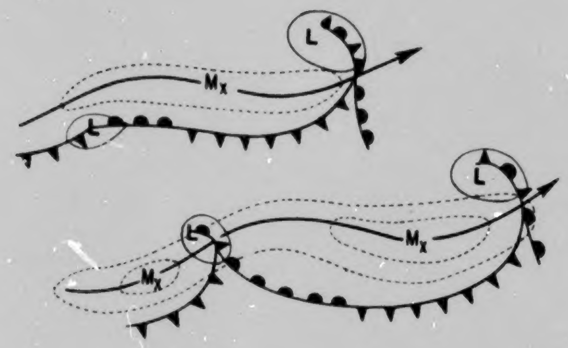

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Figure 4-22.—Usual position of surface lows in relation to moving jet maximums.

5 Although not every jetstream maximum will have an associated sea level low, each low embedded in the westerlies will be associated with a jet maximum. Lows embedded in the westerlies are those migratory lows that are under influence of the basic westerly flow.

6. The jetstream aloft should parallel the direction of the warm sector isobars of a surface low, since these isobars are alined along the upper level flow.

7. The jetstream aloft will roughly parallel the isobars around the southern periphery of a cold (slow moving) surface low.

8. The jetstream will roughly parallel the isobars along the northern periphery of a warm (slow moving) surface high.

9. When a cold, moving, polarhigh stagnates and begins to warm up, it can alter quite markedly the jetstream configuration to its rear. As the thermal contrast in the southwest quadrant is destroyed and warm air advected northward, the original jetstream dissipates and a new jet forms to the north. This is explained by the fact that the low or trough aloft associated with the surface cold high dissipates and the whole system becomes a warm cored high.

In summary, the polar front jetstream is associated with surface fronts and pressure systems in a very simple way. The jetstream wave associated with the surface low increases in amplitude as the cyclone deepens. At first the jetstream lies north of a young cyclone at the surface, moves southward, and in the latter stages of development is found south of these deep surface cyclones. At first, the jetstream does not intersect the frontal system, but the jet does cross the occluded front in the latter stages of cyclone development.

Subtropical Jetstream

Subtropical jet streams (STJ) are persistent features of the tropical general circulation. In the northern hemisphere during winter, the STJ shows a simple broadscale current which is continuous around the world with a basic three-wave pattern of ridges and maximum wind speeds over the east coasts of Asia and North America, and the Middle East. The mean latitude is 27.5° N, ranging from 20° to 35° N. The core is located near the 200 mb level, and speeds of 150 to 200 knots are not uncommon.

In the northern hemisphere during summer, the westerly jet is not in evidence; instead, the tropical easterly jet (TEJ) is a persistent feature over extreme southern Asia and northern Africa. It extends over the layer from 200 to 100 mb in the latitude belt of 5° to 20° N; core speeds over 100 knots are often observed.

In the southern hemisphere, a subtropical jet stream exists in both summer and winter as a broadscale continuous westerly current. Its mean latitude varies from 26° S in winter to 32° S in summer. The highest average core speeds range from 140 knots in July to 70 knots in January-February. The greatest variability in location and speed occur in summer and the least in winter.

Polar Night Jetstream

The polar night jetstream has been suspected for a number of years, but because of the lack of observational data in the Arctic and Antarctic regions, the actual existence of such a jetstream has been difficult to confirm. Chapter VI, titled "The Polar Night Jet Stream" from NavWeps 50-1P-549, Jet Streams of the Atmosphere, summarizes the data available on this jetstream to date.

The polar night jetstream is said to be located high in the stratosphere in and around the Arctic and Antarctic Circles. This jetstream does not appear to be associated with tropopauses as with the jetstreams of the middle and lower latitudes, but rather with heating and cooling in the ozonosphere, which takes place over the long winters and summer peculiar to that region. Nevertheless, this jetstream core is associated with one of the features of the other jetstreams in that this core is situated at an altitude where the stability of the atmosphere increases discontinuously upward. The principal research and study on this stream has been for winter.

CONVERGENCE AND DIVERGENCE

You are, no doubt, familiar with the two terms "convergence" and "divergence," as they have come to your attention when used in relation to surface lows and highs. In this section, the terms are redefined, and the motions involved in convergence and divergence, and the relationship of these processes to other meteorological processes are covered.

CONVERGENCE AND DIVERGENCE (SIMPLE MOTIONS)

Convergence

Convergence is defined as the increase of mass within a given layer of the atmosphere. In order for this to take place, the winds must be such as to result in a net inflow of air into that layer. We generally associate this type of convergence with low-pressure areas where convergence of winds toward the center of the low results in an increase of mass into the low and an upward motion. In meteorology, we distinguish between two types of convergence as either horizontal or vertical convergence, depending upon the axis of the flow.

Divergence

Divergence is defined as the decrease of mass within a given layer of the atmosphere. Winds in this situation are such that there is a net flow of air outward from the layer. We associate this type of divergence with high-pressure cells, where the flow of air is directed outward from the center, causing a downward motion there. Divergence, too, is classified as either horizontal or vertical.

The simplest form of convergence and divergence is the type that results from wind direction alone. Two flows of air need not be opposing each other to create convergence or divergence but may be at any angle to each other to create a net inflow of air for convergence or a net outflow for divergence.

Wind speed in relation to the wind direction is also a valuable indicator. For example, on a streamline analysis chart we can analyze both wind direction and wind speed, the field resulting from variations in wind speed along the streamlines or the convergence or divergence of the streamlines. The following are some of the combinations or variations of wind speed and streamlines.

1. In a field of parallel streamlines, if the wind speed is decreasing downstream (producing a net inflow of air for this layer) convergence is taking place. If the flow is increasing downstream (a net outflow of air for the layer), divergence is occurring.
2. In an area of uniform wind speed along the streamlines, if the streamlines fan out (diverge), divergence is occurring; if the streamlines converge, we can say that convergence is taking place.
3. Normally, the convergence and divergence components are combined. The fact that streamlines converge or diverge does not necessarily indicate convergence or divergence. We must also consider the wind speeds—whether they are increasing or decreasing downstream in relation to whether the streamlines are spreading out or coming together.
4. If, when looking downstream on the streamlines, the wind speed increases and the streamlines diverge, divergence is taking place. On the other hand, if the wind speed decreases downstream, and the streamlines come together, convergence is taking place.

There are other cases where it is difficult to tell whether divergence or convergence is taking place, such as when the speed of the wind decreases downstream and the streamlines spread apart, and when speed increases downstream and the streamlines converge. A special evaluation then must be made to determine the net inflow or outflow.

DIVERGENCE AND CONVERGENCE (COMPLEX MOTIONS)

In this discussion stress is placed on high-level convergence and divergence in relation to contour patterns downstream, and the advective patterns associated therewith. Low tropospheric advection (and also stratospheric advection) certainly plays a large role in pressure change mechanisms, but in the majority of cases it is thought to compensate the higher level mass changes produced by the unbalanced forces associated with the inertia of parcels of air aloft.

Since the term "divergence" is meant to denote depletion of mass, while convergence is meant to denote accumulation of mass, the prognostic analyst is concerned with the MASS divergence or convergence in estimating pressure or height changes. Mass divergence in the entire column of air produces pressure or height falls, while convergence in the entire column of air produces pressure or height rises at the base of the column.

Mass divergence and convergence involve the density field as well as the velocity field. However, the mass divergence and convergence of the atmosphere are believed to be largely stratified into two layers as follows:

1. Below about 600 millibars, velocity divergence and convergence occur chiefly in the friction layer, which is about one-eighth of the weight of the 1,000- to 600-mb advection stratum, and may be disregarded in comparison with density transport in estimating the contribution to the pressure change by the advection stratum.
2. Above 600 millibars, mass divergence and convergence largely result from horizontal divergence and convergence of velocity. However, on occasion, stratospheric advection of density may be a modifying factor.

The stratum below about the 400-mb level may be regarded as the ADVECTION stratum; above approximately the 400-mb level, the SIGNIFICANT horizontal DIVERGENCE or CONVERGENCE stratum. Also the advection stratum may be thought of as the zone in which compensation of the dynamic effects of the upper stratum occurs.

At about 8 km the density is nearly constant in time and space. This level, which is near the 350-mb pressure surface, is called the ISOPYCNIC LEVEL, a level of constant density. This level is the location of the upper tropospheric maximum of interdiurnal pressure variation, with mass variations of opposite sign above and below this level.

Since the density at 200 millibars is only four-sevenths the density at the isopycnic level, the height change at 200 millibars would be almost twice that a 350 millibars for the same pressure change at corresponding geometric levels. Thus height changes in the lower stratosphere tend to be a maximum even though pressure changes are a maximum at the isopycnic level which is usually below the midlatitude tropopause.

The large pressure variations at the isopycnic level require TEMPERATURE variations for constancy of density. Since the density is nearly constant in space at this level, the required temperature variations must result from vertical motions. When the pressures are rising at this level, the temperature must also rise to keep the density constant. A temperature rise can be produced by descending motion. Similarly falling pressures at this level require falling temperatures to keep the density constant. Falling temperatures in the absence of advection can be produced by ascent through this level.

Thus, rising heights at the pressure level of constant density are associated with subsidence and falling heights of this surface are associated with upwelling.

Subsidence at 350 millibars can result from horizontal CONVERGENCE above this level and upwelling here would result from horizontal DIVERGENCE above this level.

Since rising heights in the upper troposphere are also known to be associated with rising of the tropopause and lower stratosphere, the maximum horizontal convergence must occur between the constant density level and the average level of the tropopause (about 250 millibars). This is due to the reversal of sign of the vertical motion between the tropopause and the isopycnic level. Thus, the level of maximum horizontal velocity convergence must be between 300 millibars and 200 millibars and is the causative mechanism for pressure or height rises in the upper air. Similarly, upper height falls are produced by horizontal velocity divergence with a maximum at the same level. The maximum divergence occurs near or slightly above the tropopause and closer to 200 millibars than to 300 millibars. It is probably more realistic therefore to define a LAYER of maximum divergence and convergence as occurring between the 300- and 200-mb pressure surfaces. This is also the layer in which the core of the jetstream is usually located. At this level the cumulative effects of the mean temperature field of the troposphere produce the sharpest horizontal contrasts in the wind field.

The chart best suited for determination of convergence and divergence is the 300-mb chart. Because of the sparsity of reports at the 300-mb level, it is frequently advantageous to determine the presence of convergence and divergence at the 500-mb level. These charts will ordinarily yield good results.

The usual distribution of divergence and convergence relative to moving pressure systems is believed to be as follows: (Also see ch. 11.)

1. In advance of the low, convergence occurs at low levels and divergence occurs aloft, with the level of nondivergence at about 600 millibars on the average.
2. In the rear of the low, there is usually convergence aloft and divergence near the surface.

The low-level convergence ahead of the low occurs usually in the stratum of strongest warm advection, and the low-level divergence in the rear of the low occurs in the stratum of strongest cold advection. The low-level divergence occurs principally in the friction layer, and it is thought to be of minor importance in modifying estimates of thickness advection compared with heating and cooling from the underlying surfaces.

In advance of the low, the air rises in response to the low-level convergence with the maximum ascending motion at the level of nondivergence eventually becoming zero at the level of maximum horizontal divergence. Above this level, descending motion is believed to set in. In the rear of the low the reverse is true; that is, descending motion in the surface divergent stratum and ascending motion in the upper troposphere above the level of maximum horizontal convergence. In deepening systems the convergence aloft in the rear of the low is small or may even be negative (divergence). In filling systems the divergence aloft in advance of the low is small or even negative (convergence).

Thus, in the travel and development of highs and lows, two vertical circulations are involved, one below and one above the 300-mb level approximately. The LOWER VERTICAL CIRCULATION is upward in the cyclone, thence toward the anticyclone, and then downward in the anticyclone. The UPPER VERTICAL CIRCULATION involves subsidence in the stratosphere of the developing cyclone and ascent in the upper troposphere and lower stratosphere of the developing anticyclone. (See fig. 4-23.)

Divergence and upper-height falls are associated with high-speed winds approaching weak contour gradients which are cyclonically curved. Figure 4-24 illustrates schematically some contour patterns associated with contour falls.

Convergence and upper-height rises are associated with:

1. Low-speed winds approaching straight or cyclonically curved strong contour gradients.
2. High-speed winds approaching anticyclonically curved weak contour gradients.

Figure 4-25 illustrates schematically the contour patterns associated with each of these cases of contour rises. The height rises and falls occur downstream and to the left of the flow.

The technique for determining the areas of divergence consists in noting the areas where winds of high speed are approaching weaker pressure gradients of straight or cyclonic curvature downstream. When inertia carries a high-speed parcel of air into a region of weak pressure gradient, it possesses a Coriolis force too large to be balanced by the weaker pressure gradient force. It is thus deflected to the right by the resultant unbalanced force to the right. This results in a deficit of mass to the left due to the

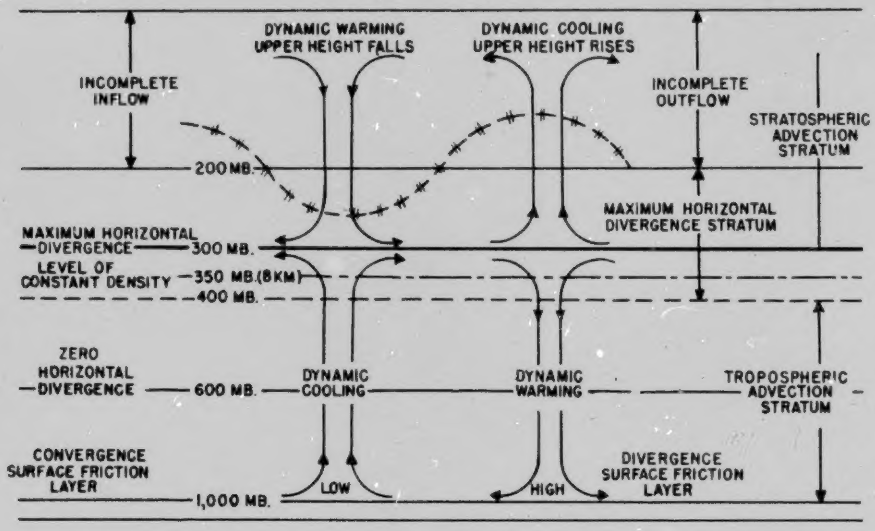
DYNAMIC WARMING
UPPER HEIGHT FALLS
DYNAMIC COOLING
UPPER HEIGHT RISES
INCOMPLETE
INFLOW
INCOMPLETE
OUTFLOW
STRATOSPHERIC
ADVECTION
STRATUM
200 MB.
MAXIMUM HORIZONTAL
DIVERGENCE STRATUM
MAXIMUM HORIZONTAL
DIVERGENCE
300 MB.
LEVEL OF
CONSTANT DENSITY
350 MB.(8KM)
400 MB.
ZERO
HORIZONTAL
DIVERGENCE
600 MB.
DYNAMIC
COOLING
DYNAMIC
WARMING
TROPOSPHERIC
ADVECTION
STRATUM
CONVERGENCE
SURFACE FRICTION
LAYER
1,000 MB.
LOW
HIGH
DIVERGENCE
SURFACE FRICTION
LAYER

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Figure 4-23.—Generalized vertical circulation over developing highs and lows.

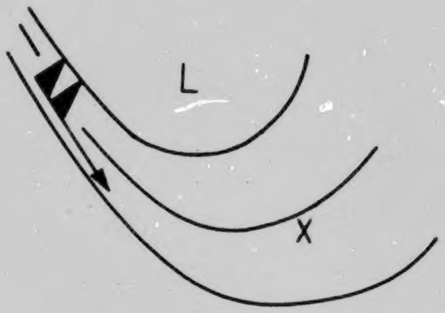

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Figure 4-24.—Divergence illustrated.

failure of incoming parcels to compensate the longitudinal deficit of mass by the parcels leaving the particular area. The parcels which are deflected to the right must penetrate high pressure and are thus slowed down until they are in balance with the weaker pressure gradient. Then they can be steered along the existing isobaric or contour channels.

If the weak gradients downstream are cyclonically curved, the divergence resulting from the influx of high-speed wind is even more marked due to the additional effect of centrifugal forces (on the other hand, the effect of centrifugal forces of anticyclonically curving high-speed parcels is of extreme importance in producing overshooting of high-speed air from sharply curved ridges into adjacent troughs causing pressure rises in the west side of the troughs.)

If the influx of high-speed parcels toward diverging cyclonically curved contours is sustained, large contour falls will occur downstream to the left of the inertial path of the strong winds. Eventually a strong pressure gradient is propagated downstream to the right of the high-speed winds, chiefly by pressure falls to the left of the direction of high-speed winds in the cyclonically curved contours with weak pressure gradient. Usually the deflection of air toward higher pressure is so slight that it is hardly observable in individual wind observations. However, when the pressure field is very flat to the right of the incoming high-speed stream, noticeable angles between the wind and contours may be observed, especially at lower levels, due to transport of momentum downward as a result of subsidence, where the gradients are even weaker. This occurs sometimes to such an extent that

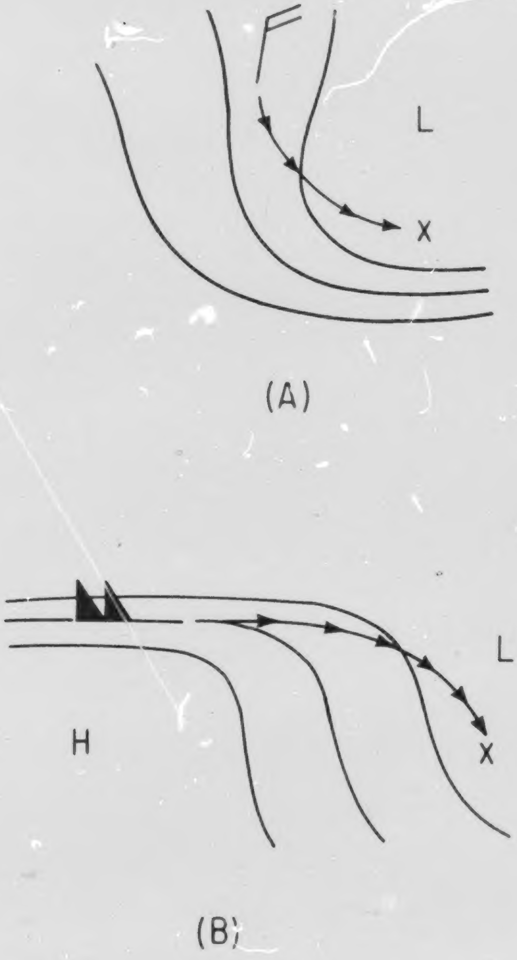

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Figure 4-25.—Convergence illustrated.

the streamlines are considerably more curved anticyclonically than the contours. In rare cases this results in anticyclonic circulation centers out of phase with the high-pressure center. This is, of course a transitory condition necessitating a migration of the pressure center toward the circulation center, and this is of prognostic value. In cases where the high-pressure center and anticyclonic streamline center are out of phase, the pressure center will migrate toward the circulation center (which is usually a center of mass convergence).

It is more usual, however, for the wind component toward high pressure to be very slight, and unless the winds and contours are drawn with great precision, the deviation goes unnoticed.

High-speed winds approaching sharply curved ridges result in large height rises downstream from the ridge due to OVERSHOOTING of the highspeed air. As is well known from the gradient wind equation, for a given pressure gradient there is a limiting curvature to the trajectory of a parcel of air moving at a given speed. Frequently on the upper air charts, sharply curved stationary ridges are observed with winds of high speed approaching the ridge. The existence of a sharply curved extensive ridge usually means a well-developed trough not far downstream, and frequently a cold or cutoff low exists in this trough. The high-speed parcels approaching the ridge, because of centrifugal forces, are unable to make the sharp turn necessary to follow the contours. These parcels overshoot the ridge anticyclonically, but with less curvature than the contours, resulting in their plunging across contours toward low pressure downstream from the ridge, and their accelerating. This may result in any one of a number of consequences for the trough downstream, depending on the initial configuration of the ridge and trough, but all are based on the convergence of mass into the trough as a result of overshooting of air from the ridge. Four of the most frequent consequences may be enumerated as follows:

1. Filling of the portion of the trough downstream from the ridge. This happens if the contour gradient is strong on the east side of the trough; that is, a blocking ridge to the east of the trough.
2. Accelerating the cutoff low out of its stationary position. This usually occurs in all cases.
3. Radically reorienting the trough. This usually happens where the trough is initially NE-SW, resulting in a N-S and in some cases a NW-SE orientation after sufficient time (36 hours).
4. It may actually cut off a low in the lower end of the trough. This usually happens when the high-speed winds approaching the ridge are

southwesterly and approach the ridge at a comparatively high latitude relative to the trough. This frequently reorients the trough line towards a more NE-SW direction. Usually, the reorientation of the trough occurs simultaneously with 1 and 2.

Closely related to the above situation are cases of sharply curved ridges where the pressure gradient in the sharply curved portion (usually the northern portions of a north-south ridge) has momentarily built up to a strength that is incompatible with the anticyclonic curvature. Such ridges often collapse with great rapidity subsequent to the development of such excessive gradients, causing rapid filling of the adjacent trough downstream and large upper contour falls where the ridge now exists. The gradient wind relation implies that subsequent trajectories of the high-speed parcels generated in the strong ridge line gradient must be less anticyclonically curved than the contours in the ridge.

It can be shown from the gradient wind equation that the anticyclonic curvature increases as the difference between the actual wind and the geostrophic wind increases, until the actual wind is twice the geostrophic, when the trajectory curvature is a maximum. This fact can be utilized in determining the trajectory of high-speed parcels approaching sharply curved stationary ridges, or sharply curved stationary ridges in which the contour gradient is excessive. By measuring the geostrophic wind in the ridge, the maximum trajectory curvature can be obtained from the gradient wind scale. This trajectory curve is the one which an air parcel at the origin point of the scale will follow until it intersects the correction curve from the geostrophic speed to the displacement curve of twice the geostrophic speed.

If actual wind speed observations are available for parcels approaching the ridge, comparison can be made with the geostrophic winds (pressure gradient) in the ridge. If the actual speeds are more than twice the measured geostrophic wind in the ridge, the anticyclonic curvature of these high-speed parcels will be less than the MAXIMUM TRAJECTORY curvature obtained from the gradient wind scale, and even greater overshooting of these high-speed parcels will occur across lower contours. Convergence in the west side of the downstream trough results in lifting of the tropopause with dynamic cooling, and upper-level contour rises.

Low-speed parcels of air from a weak pressure gradient which enter an area of stronger gradient become subject to an unbalanced gradient force toward the left due to the weaker Coriolis force. These sub-gradient winds are deflected toward lower pressure, crossing contours and producing contour rises in the area of cross-contour flow. This cross-contour flow accelerates the air until it is moving fast enough to be balanced by the stronger pressure gradient. Due to the acceleration of the slower oncoming parcels of air, the contour rises propagate much faster than might be expected on the basis of the slow speed of the air as it initially enters the stronger pressure gradient.

The following two rules summarize the above discussion:

1. High-speed winds approaching low-speed winds with weak cyclonically curved contour gradients are indicative of divergence and upper-height falls downstream and to the left of the current.
2. Low-speed winds approaching high-speed winds with strong cyclonically curved contour gradients or high-speed winds approaching low-speed winds with weak anticyclonically curved contour gradients are indicative of convergence and upper height rises downstream and to the left and right of the current, respectively.

IMPORTANCE OF CONVERGENCE AND DIVERGENCE

Convergence and divergence have a pronounced effect upon the weather occurring in the atmosphere. Vertical motion, either upward or downward, is recognized as an important parameter in the atmosphere. For instance, extensive regions of precipitation associated with extratropical cyclones are regions of large-scale upward motion of air through most of the troposphere. Similarly, the nearly cloud-free regions in large anticyclones are regions in which air is subsiding through a large portion of the troposphere. Vertical motions also affect temperature, humidity, and other meteorological variables either on a small scale (locally) or on a large scale as pointed out above.

Changes in Stability

When convergence or divergence occurs, whether on a large or small scale, it can have a very pronounced effect on the stability of the air. For example, when convection is induced by convergence, air is forced to rise without the addition of heat. If this air is unsaturated, it cools first at the dry adiabatic rate; or if saturated, at the moist rate. The end result is that the air is cooled, which will increase the instability of that air column by increasing the lapse rate. Clouds and weather often result from this process.

Conversely if air subsides, and this process is produced by convergence or divergence, the air sinking will heat at the dry adiabatic lapse rate. The warming at the top of an air column will increase the stability of that air column by reducing the lapse rate. Such warming often dissipates existing clouds or prevents the formation of new clouds. When sufficient warming due to the downward motion takes place, a subsidence inversion is produced.

Effect on Weather

The full aspects of the relationship of vertical motion to weather are explored in Vertical Motion and Weather, NWRF 30-0359-024. The most improtant application of vertical motion is the prediction of rainfall probability and rainfall amount. In addition, vertical moiton affects practically all meteorological properties, such as temperature, humidity, wind distribution, and particularly stability. In the following section the distribution of large-scale and small-scale vertical motions are considered.

Since cold air has a tendency to sink, subsidence is likely to be found to the west of upper tropospheric troughs, and rising air to the east of the troughs. Thus, there is a good relation between upper air meridional flow and vertical flow.

In the neighborhood of a straight Northern Hemisphere jetstream, convergence is found to the north of the stream behind centers of maximum speed as well as to the south and ahead of such centers. Divergence exists in the other two quadrants. Below the regions of divergence the air rises; below those of convergence there is subsidence.

These general rules of thumb are not perfect, and only yield a very crude idea about distribution of vertical motion in the horizontal. Particularly over land in summer, there exists little relation between large-scale weather patterns and vertical motion. Rather, vertical motion is influenced by local features and shows strong diurnal variations, the phase which may be quite different in different regions. Large-scale vertical motion is of small magnitude at the ground (zero if the ground is flat). Above the ground, it increase in absolute magnitude to at least 500 millibars and decreases in the neighborhood of the tropopause. Reversals in sign in a vertical column are rare in the troposphere, but are likely to occur in the stratosphere. There have been several studies of the relation between frontal precipitation and large-scale vertical velocities, computed by various techniques. In all cases, the probability of precipitation is considerably higher in the 6 hours following an updraft than following subsidence. Clear skies are most likely with downdrafts. On the other hand, it is not obvious that large-scale vertical motion is related to showers and thunderstorms caused during the daytime by heating. However, squall lines, which are formed along lines of horizontal convergence show that large-scale vertical motion may also play an important part in convective precipitation.

Vertical velocity charts are currently being transmitted over the facsimile network by the National Meteorological Center and are computed by Numerical Weather Prediction methods. The charts have plus signs indicating upward motion and minus signs indicating downward motion. The figures indicate vertical velocity in centimeters per second (cm/sec). With the larger values of upward motions (plus values) the likelihood of clouds and precipitation increases. However, an evaluation of the moisture and vertical velocity should be made to get optimum results. Obviously, upward motion in dry air is not as likely to produce precipitation as upward motion in moist air.

ROTATIONAL MOTION AS IT AFFECTS THE ATMOSPHERE

Many forecasters and meteorologists started their careers years ago when the main concern was with small areas and short-time intervals. The geographical extension of observations, especially those of the upper air, and the study of these observations by dynamic meteorologists have proved (although it had long been suspected) that the cyclone or anticyclone is not independent of developments in other parts of the hemisphere; rather it constitutes a cogwheel in a larger mechanism. The relationship of the cyclone to the larger-scale flow patterns must therefore be a part of the daily forecast routine if progress is to be made in forecasting the motions of cyclones and associated upper wind fields.

One of the studies that has gained prominence in recent years has been that of vorticity. Unfortunately many forecasters have a tendency to shy away from the subject of vorticity, as they consider it too deep a subject to be mastered. This means they are neglecting an important forecasting tool. The principles of vorticity are no more complicated than most of the principles of physics, and can be understood just as easily. During the last few years vorticity has become a prominent tool in meteorology, especially in the field of Numerical Weather Prediction. For this purpose, we will consider in this section of the chapter the meaning of vorticity, its evaluation, and its relationship to other meteorological parameters.

VORTICITY

Vorticity measures the rotation of very small air parcels. A parcel has vorticity when the parcel spins as it moves along its path. On the other hand, a parcel which does not spin is said to have zero vorticity. The axis of spinning or rotation can extend in any direction, but for our purposes, we are mainly concerned with the rotational motion about an axis that is perpendicular to the surface of the earth. For example, we could drop a chip of wood into a creek and watch its progress. The chip will move downstream with the flow of water but it may or may not spin as it moves downstream. If it does spin, the chip has vorticity. When we try to isolate the cause of the spin, we find that two properties of the flow of water cause the chip to spin: (1) If the flow of water is moving faster on one side of the chip than the other, this is shear of the current; (2) if the creek bed curves, the path has curvature. Vorticity always applies to extremely small air parcels; thus, a point on one of our upper air charts may represent such a parcel. We can examine this point and say that the parcel does or does not have vorticity. However, for this discussion, larger parcels will have to be used to more easily visualize the effects. Actually, a parcel in the atmosphere has three rotational motions at the same time: (1) Rotation of the parcel about its own axis (shear), (2) rotation of the parcel about the axis of a pressure system (curvature), and (3) rotation of the parcel due to the atmospheric rotation. The sum of the first two components is known as relative vorticity, and the sum total of all three is known as absolute vorticity.

Relative Vorticity

Relative vorticity is the sum of the rotation of the parcel about the axis of the pressure system (curvature) and the rotation of the parcel about its own axis (shear). The vorticity of a horizontal current can be broken down into two components, one due to curvature of the streamlines and the other due to shear in the current.

SHEAR.—First, let us examine the shear effect by looking at small air parcels in an upper air pattern of straight contours. Here the wind shear results in each of the three parcels having different rotations. (See fig. 4-26.)

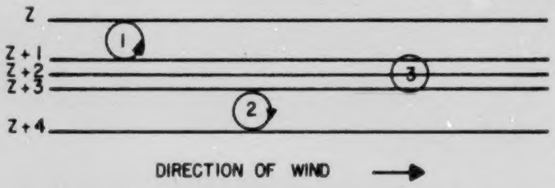

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Figure 4-26.—Illustration of vorticity due to the shear effect.

1. Parcel No. 1 has stronger wind speeds to its right. As the parcel moves along, it will be rotated in a counterclockwise direction.
2. Parcel No. 2 has the stronger speed to its left; therefore, it will rotate in a clockwise direction as it moves along.
3. Parcel No. 3 has speeds evenly distributed. This parcel has equal speeds at opposite ends of the diameter. It will move, but it will not rotate. It is said to have zero vorticity.

Therefore, to briefly review the effect of shear—a parcel of the atmosphere has vorticity (rotation) when the wind speed is stronger on one side of the parcel than on the other.

Now let's define positive and negative vorticity in terms of clockwise and counterclockwise rotation of a parcel. The vorticity is positive when the parcel has a counterclockwise rotation and the vorticity is negative when the parcel has clockwise rotation (anticyclonic, Northern Hemisphere).

Thus, in figure 4-26, parcel No. 1 has positive vorticity, and parcel No. 2 has negative vorticity.

CURVATURE.—Vorticity can also result from curvature of the airflow or path. In the case of the wood chip flowing with the stream, the chip will spin or rotate as it moves along if the creek curves.

To demonstrate the effect of curvature, let us consider a pattern of contours having curvature but no shear. (See fig. 4-27.)

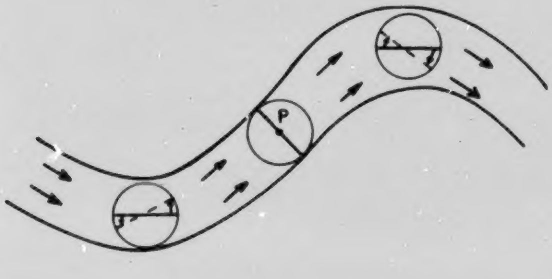

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Figure 4-27.—Illustration of vorticity due to curvature effect.

Place a small parcel at the trough and ridge lines and observe the way in which the flow will spin the parcel, causing vorticity. The diameter of the parcel will be rotated from the solid line to the dotted position (due to the northerly and southerly components of the flow on either side of the trough and ridge lines).

Note that we have counterclockwise rotation at the trough (positive vorticity), and at the ridge line we have clockwise rotation (negative vorticity). At the point where there is no curvature (inflection point), there is no turning of the parcel, hence no vorticity. This is demonstrated at point P in figure 4-27).

COMBINED EFFECTS.—To find the relative vorticity of a given parcel we must consider both the shear and curvature effects. It is quite possible to have two effects counteract each other; that is, where shear indicates positive vorticity but curvature indicates negative vorticity, or vice versa. (See fig. 4-28.)

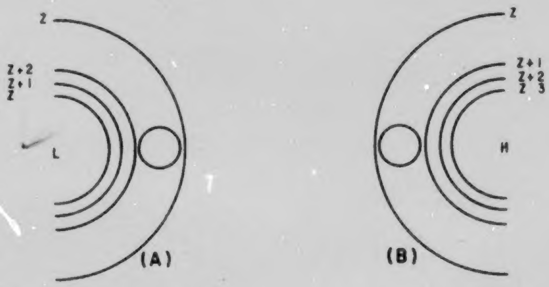

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Figure 4-28.—Illustration of shear effect opposing the curvature effect in producing vorticity. (A) Negative shear and positive curvature; (B) positive shear and negative curvature.

To find the net result of the two effects we would measure the value of each and add them algebraically. The measurement of vorticity will be discussed in the next section.

It must be emphasized here that relative vorticity is observed instantaneously. It is not necessary for a parcel to rotate a large amount to have vorticity. In summary, relative vorticity in the atmosphere is defined as the instantaneous rotation of very small particles. The rotation results from wind shear and curvature. We refer to this vorticity as being relative, because all the motion illustrated was relative to the surface of the earth with latitude and longitude lines used as reference points.

MEASUREMENT OF RELATIVE VORTICITY.—The various mathematical expressions for vorticity are merely "shorthand" ways of saying that vorticity may be determined by shear and curvature. One common expression for relative vorticity is:

$$Zr = \frac{v}{R} - \frac{\Delta v}{\Delta n}$$

where:

Zr indicates vorticity of the wind field at a constant level or constant pressure surface (the so-called vertical component of vorticity). This is expressed in radians/sec.

v is the wind speed, in terms of knots or meters/second, and represents the speed of movement of the air parcel.

R is the radius of the windflow. R is positive for cyclonic curvature and negative for anticyclonic curvature.

Δv represents the change in speed of the airflow along a line perpendicular to the contours.

Δn is the direction along which we measure the shear vorticity and represents the distance between the two points for which v was determined.

To illustrate how this equation is evaluated to find geostrophic vorticity, a sample contour field having both shear and curvature is presented in figure 4-29.

To find the vorticity at P, proceed in exactly the same manner regardless of where P is located in the pattern. It must be noted, however, that R is positive for cyclonic curvature and negative for anticyclonic curvature, and the Δn is perpendicular to the flow and measured from high to low contours, or from left to right when looking downstream.

In the example, figure 4-29, R may be determined from a template. The geostrophic wind at P gives the value of v. These two values give the curvature vorticity, v/R. If 5 is 500,000 meters and v is 50 meters/sec, the

$$\frac{v}{R} = \frac{50\ \text{m/sec}}{500{,}000\ \text{m}} = \frac{1}{10{,}000\ \text{sec}} = 10^{-4}\ \text{sec}^{-1}$$

To investigate the shear term

$$\frac{\Delta v}{\Delta n} = - \frac{v_2 - v_1}{\Delta n}$$

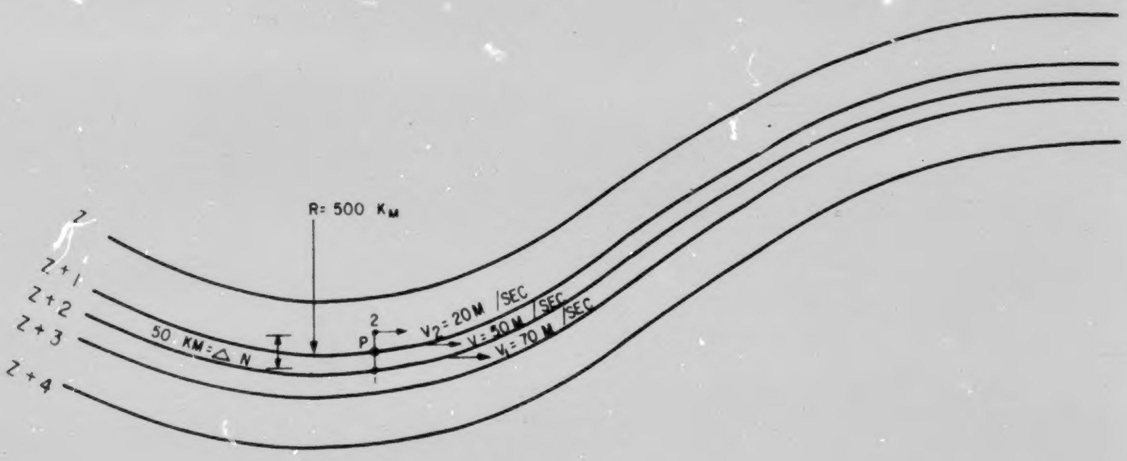

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Figure 4-29.—Evaluation of vorticity using the vorticity equation.

at point P, we measure the wind speed at points v_1 and v_2, and Δn.

$$v_2 = 20 \text{ m/sec}$$
$$v_1 = 70 \text{ m/sec}$$
$$\Delta n = 50{,}000 \text{ m}$$

Therefore,

$$-\frac{\Delta v}{\Delta n} = -\frac{v_2 - v_1}{\Delta n} = -\frac{20\text{-}70 \text{ m/sec}}{50{,}000 \text{ m}}$$

$$= \frac{1}{10{,}000 \text{ sec}}$$

Thus, the sum of the shear and curvature vorticities for the selected point P is:

$$Zr = \frac{v}{r} - \frac{\Delta v}{\Delta n} = \frac{1}{10{,}000 \text{ sec}} + \frac{1}{10{,}000 \text{ sec}}$$

$$= \frac{2}{10{,}000 \text{ sec}} \text{ or } 2 \times 10^{-4} \text{ sec}^{-1}$$

Vorticity is usually expressed as radians/sec. Radians are used to measure angles in the same manner as degrees. Whereas a circle has 360°, it has 6.28 radians; so a radian is equal to about 57°. These units, radians/sec, demonstrate the idea of vorticity as an instantaneous rotational speed, so that for radians/sec we merely write 1/sec or sec^{-1}. There are numerous other ways that vorticity can be measured. This is only one example.

In summary, we must keep in mind that cyclonic vorticity is positive and anticyclonic vorticity is negative. Therefore, when vorticity is said to increase, it is becoming more cyclonic and less anticyclonic. When it decreases, the converse is true.

Absolute Vorticity

When the relative vorticity of a parcel of air is observed by a person completely removed from the earth, he observes an additional component of vorticity created by the rotation of the earth. Thus, this person sees the total or absolute vorticity of the same parcel of air.

The total vorticity, that is, relative vorticity (Zr) plus that due to the earth's rotation, is known as the ABSOLUTE VORTICITY. As was stated before, for practical use in meteorology, only the vorticity about an axis perpendicular to the surface of the earth is considered. In this case, the vorticity due to the earth's rotation becomes equal to the CORIOLIS PARAMETER. This is expressed as $2\omega \sin \phi$, where ω is the angular velocity of the earth and ϕ is the latitude. Therefore, the absolute vorticity is equal to the Coriolis parameter plus the relative vorticity. Writing this in equation forms gives: (Za = absolute vorticity)

$$Za = 2\,\omega \sin \phi + Zr$$

Current and prognostic charts of absolute vorticity are currently prepared by Numerical Weather Prediction methods and transmitted by the National Meteorological Center on the facsimile network.

CONSTANT ABSOLUTE VORTICITY TRAJECTORIES (CAVT)

It has been proved mathematically (under some rigid assumptions) that parcels moving in the atmosphere conserve their absolute vorticity as they move from place to place; that is, the value of Za does not change. The most important of the assumptions is that no convergence or divergence occurs. Actual observations under conditions of no convergence or divergence bear out this theory.

Absolute vorticity remaining constant, the motion of a parcel of air as it moves from latitude to latitude in the Northern Hemisphere will describe a definite pattern called Constant Absolute Vorticity Trajectories (CAVT's). Let us now consider the movement of some of these air parcels.

Coriolis Parameter

As a parcel moves from south to north, the latitude increases; there the CORIOLIS PARAMETER increases. If the value of $2\,\omega \sin \phi$ increases, the value of Zr must decrease in order that the sum remains the same. If Zr decreases, it becomes less cyclonic or, in other words, more anticyclonic. This being the case, the parcel of air would move in a path which becomes less and less cyclonic with increasing latitude, and eventually becomes anticyclonic. Then the anticyclonic curvature increases until the parcel is moving from west to east. Continuing to move in an anticyclonic path, the parcel begins to turn back toward the south. (See fig. 4-30.)

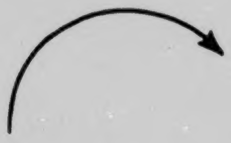

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Figure 4-30.—Coriolis parameter increasing.

As a parcel moves from north to south, the latitude decreases, and hence the Coriolis parameter decreases. If the value of 2 ω sin ϕ decreases, the value of Zr must increase in order that their sum remain the same. If Zr increases, it becomes more cyclonic or less anticyclonic. In other words, the parcel of air would move in a path which becomes less and less anticyclonic, and eventually becomes cyclonic. Then the cyclonic curvature increases up to a critical latitude where the parcel will move from west to east and begin to turn back toward the north. (See fig. 4-31.)

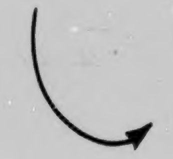

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Figure 4-31.—Coriolis parameter decreasing.

With these forces and no others acting on the parcel, it is evident that the parcel would describe a sinusoidal path as it moves through space. This is illustrated in figure 4-32.

The points A, B, C, and D, where the curvature changes from cyclonic to anticyclonic, or vice versa, are known as inflection points.

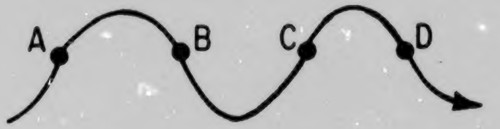

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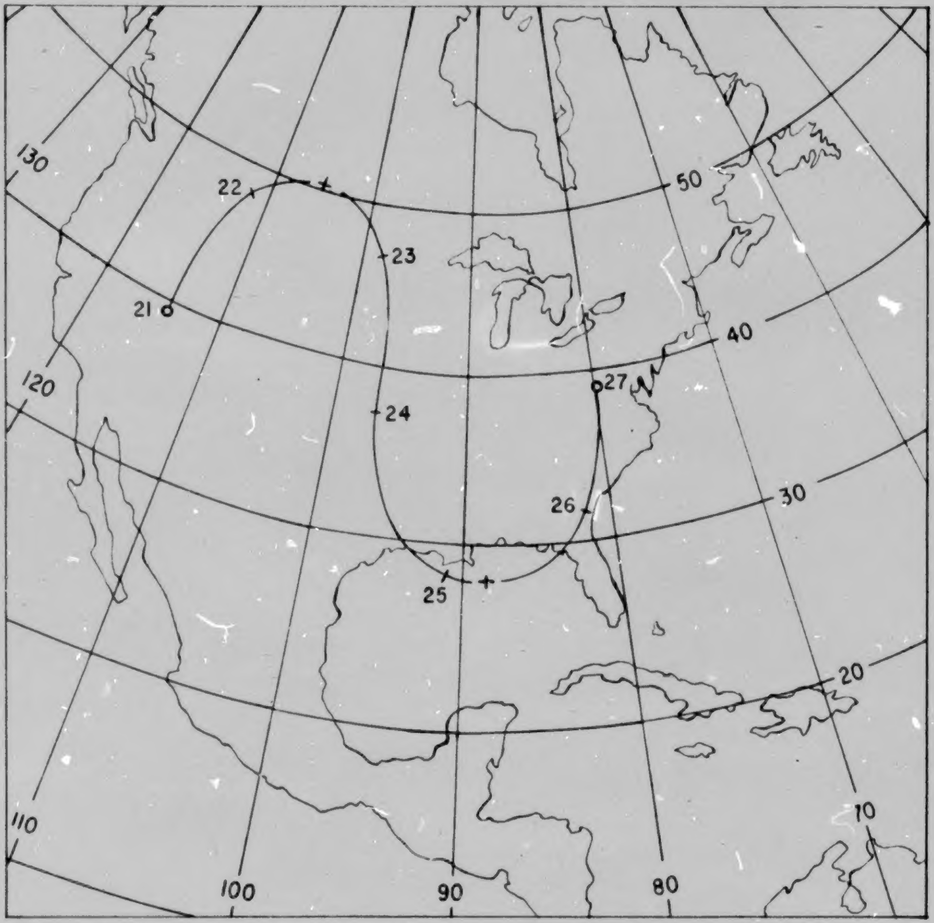
Figure 4-32.—Sinusoidal vorticity path.

Amplitude and Length

The amplitude and length of the waves depend upon the initial speed and direction of the parcel and its initial latitude. This fact is fundamental in determining constant vorticity trajectories. If the speed and direction of a parcel can be determined at the inflection point and the latitude of the inflection point is known, the future path of the parcel can be determined. When the inflection angle (the angle at the inflection point between the latitude circle and the direction of motion of the parcel) is small, the waves tend to be flat and elongated. Both the amplitude and the wavelength increase with an increase in speed. For values of the inflection angle greater than 90°, but less than 130°, from east, the path becomes a figure eight.

In figure 4-33 the parcel of air starts out at 39° N lat. with a speed of 25 knots, with an inflection angle of +90°. (Heading: due north.)

The Coriolis parameter increases, and relative vorticity (Zr) decreases in order that 2 ω sin ϕ + Zr = Za. The parcel begins to turn slowly anticyclonically until it reaches its critical latitude. The ciritical latitude of a parcel of air with a certain speed and inflection angle, starting from a certain latitude, is that latitude at which it has reached maximum amplitude in its sinusoidal wave pattern, and must now, due to the forces acting upon it, reverse its direction of travel. All parcels with different initial inflection angles, speed, and latitude have their own critical latitude, that latitude being dependent upon all three factors: The inflection angle, speed, and latitude of the parcel. The parcel of air in figure 4-35 reached its critical latitude at



SAMPLE C A V TRAJECTORY. (STROKES AND NUMERALS ALONG THE PATHS MARK POSITIONS AND DATES FOR 24-HOUR INTERVALS FROM THE STARTING POINT ON THE 2300 EST MAP OF THE INITIAL DATE.)

AG.444

Figure 4-33.—Sample CAV trajectory.

50° N. We see the parcel heading eastward, and in so doing, the Coriolis parameter decreases, the relative vorticity increases, and the parcel then heads southward in an ever lessening anticyclonic path until the path is straight. Inasmuch as the Coriolis parameter is still decreasing, relative vorticity is still increasing, and the parcel begins to travel in a cyclonic path. When the parcel was in straightline flow at 39° N lat, it was midway between a cyclonic and anticyclonic curvature. This point is referred to as an inflection point and will later be an important factor in forecasting upper air long-wave movements.

The parcel continues its southward movement, increasing its relative vorticity because the Coriolis parameter is decreasing. At 28° N lat., the parcel has again reached its critical latitude. The Coriolis parameter now begins to increase again and the parcel flows northward in a path that is becoming less and less cyclonic, until, once again, at 39° N lat., it has reached its inflection point, having been displaced in this process over 2,000 miles to the east. Without other considerations entering the picture, this process would then repeat itself interminably.

In the above discussion, it was assumed that no horizontal convergence occurs. However, parcels depart from the CAV trajectories in regions of horizontal convergence and divergence. If a column of air, as shown in figure 4-34 of height D is considered and it is assumed that no change in the volumn occurs, it is found that the height increase with horizontal convergence and decreases with horizontal divergence.

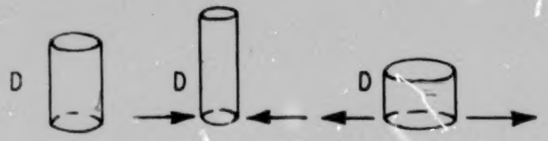


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Figure 4-34.—Height changes with horizontal convergence and divergence.

VORTICITY THEOREM

It can be shown mathematically that relative vorticity is related to horizontal divergence and convergence in the following manner:

$$\frac{2\,\omega \sin \phi + Zr}{D} = \text{Constant}$$

This is known as the VORTICITY THEOREM and shows that the sum of the Coriolis parameter and the relative vorticity of the parcel divided by the height of the column must equal a constant value. This means that if one of the values changes, one or both of the others must change so that the quotient will remain the same.

How does the vorticity theorem explain the motion of a parcel of air as the parcel undergoes horizontal divergence or convergence and as it moves from latitude to latitude?

Latitude

If the vorticity theorem is used, the parcel's relative vorticity must decrease if the latitude remains unchanged and horizontal divergence takes place. This is because D decreases with horizontal divergence; therefore, Zr must decrease in order that $2\,\omega \sin \phi$ plus Zr/D will equal the same value as Zr did before the decrease in D. When Zr decreases, it becomes less cyclonic or more anticyclonic, and the parcel moves in a more anticyclonic path than it would if no divergence occurs.

If the latitude remains unchanged and horizontal convergence takes place, the parcel's relative vorticity must increase. This is because D increases with horizontal convergence; therefore, Zr must increase in order that the conditions of the vorticity theorem may be satisfied. When Zr increases, it becomes more cyclonic or less anticyclonic, and the parcel moves in a more cyclonic path than it would if no convergence occurs.

Southward Movement

If a parcel moves southward, there is a decrease in $2\,\omega \sin \phi$, which requires that Zr increase if D remains the same, in order that $2\,\omega \sin \phi$ plus Zr/D remain constant. However, if horizontal divergence is taking place, D is becoming less, so that with horizontal divergence and southward motion, the denominator and one term in the numerator of the vorticity theorem are decreasing. Consequently, the decrease of these two values must be such that $2\,\omega \sin \phi$ plus Zr/D remains constant, or there must be a compensating decrease in Zr. The latter is the normal case since, as explained in the previous two paragraphs, it has been shown that Zr and D increase or decrease together at a given latitude. This fact also eliminates the case where there might be such a large decrease in $2\,\omega \sin \phi$ and such a small decrease in D that Zr must increase. A decrease in the relative vorticity of the parcel results in its moving more anticyclonically. In this case, the parcel would follow the dashed line as shown in figure 4-35 rather than the solid line (assuming that the divergence is taking place in the vicinity of the point of inflection). The solid line represents the path if no convergence or divergence occurs.

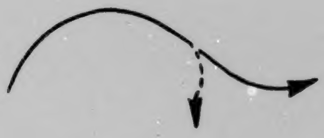

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Figure 4-35.—Parcel moving southward.

If horizontal convergence is taking place while a parcel moves southward, there is a decrease in 2 ω sin ϕ and an increase in D. Consequently, there must be a compensating increase in Zr. Therefore, the parcel acquires cyclonic vorticity more rapidly than if no convergence were occurring and curves eastward more rapidly. This is illustrated in figure 4-36 by the dashed line.

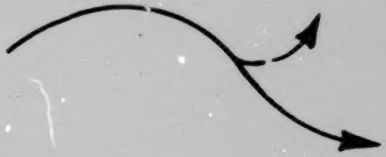

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Figure 4-36.—Horizontal convergence while a parcel moves southward.

From the above deductions, it is obvious that when parcels moving southward curve more anticyclonically than their CAV trajectories, horizontal divergence is occurring; when horizontal divergence is expected in a current moving from north to south, the path of the parcel will be more anticyclonic than expected from the CAV trajectory. Also, when parcels move southward and curve more cyclonically than their CAV trajectories, horizontal convergence is occurring; when horizontal convergence is expected in a current moving from north to south, the path of the parcel will be more cyclonic than expected from the CAV trajectory.

Northward Movement

If a parcel moves northward, there is an increase in 2 ω sin ϕ which requires that Zr decrease if D remains the same. However, if horizontal convergence is taking place, D is becoming larger. With horizontal convergence and northward motion, the denominator and one term in the numerator of the vorticity theorem are increasing. Consequently, the increase of these two values must be such that

$$\frac{2\ \omega \sin \phi \text{ plus Zr}}{\text{D}}$$

remains constant, or there must be a compensating increase in Zr. The latter is the normal case and results in the parcel moving more cyclonically than it would if there were no convergence. In this case, the parcel would follow the dashed line as shown in figure 4-37 rather than the solid line.

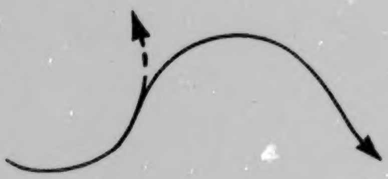

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Figure 4-37.—Parcel moving northward.

If horizontal divergence is taking place while a parcel moves northward, there is an increase in 2 ω sin ϕ and a decrease in D. Consequently, there must be a compensating decrease in Zr. Therefore, the parcel acquires anticyclonic vorticity more rapidly than if no convergence were occurring and curves eastward more rapidly. This is illustrated in figure 4-38 by the dashed line.

From this discussion, it is evident that when parcels moving northward curve more cyclonically than their CAV trajectories, horizontal convergence is taking place. When horizontal convergence is expected in a current moving south to north, the path of the parcel will be more cyclonic than expected from the CAV trajectory. Also, when parcels moving northward curve more anticyclonically than their CAV trajectories, horizontal divergence is occurring.

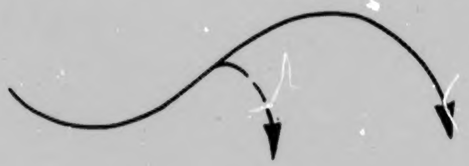

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Figure 4-38.—Horizontal divergence while a parcel moves northward.

When horizontal divergence is expected in a northward moving current, the path of the parcel will be more anticyclonic than expected from the CAV trajectory.

USE OF CAVT'S

The appendix of Practical Methods of Weather Analysis and Prognoses, NavAer 50-1P-502, contains CAVT tables for intervals of 5° of latitude. The tables are entered according to the initial latitude, and the wind speeds and wind directions of the air parcels at their inflection points. The tabular values are in one-fourth wavelengths. The results of these tables are a series of waves in which the upper and lower portions of the waves are symmetrical, though not necessarily of equal size. The primary problem in moving long waves by CAVT's is the determination of a valid inflection point. The definition of the inflection point requires straight line flow (that is, midway between the cyclonic and the anticyclonic flow) and no shear.

In the case of a jet max being embedded in the westerly flow, the CAVT's cannot be used at face value because of the convergence and divergence in the currents. It must be decided whether there will be convergence or divergence and whether the CAVT must be readjusted accordingly. If the path of the parcel is toward diverging cyclonic contours, the actual path will be to the right of the CAV path. If the path of the parcel is toward converging contours, the actual path is to the left of the CAV path. The net result of divergence is that the actual amplitude of the troughs will be greater than CAV amplitudes and trough lines will be farther east than CAV positions. For ridges, the actual contours will be flatter than CAVT's and farther east as well. The net result of convergence is that actual troughs will have less amplitude and will be located farther west than CAV troughs as shown in figure 4-39.

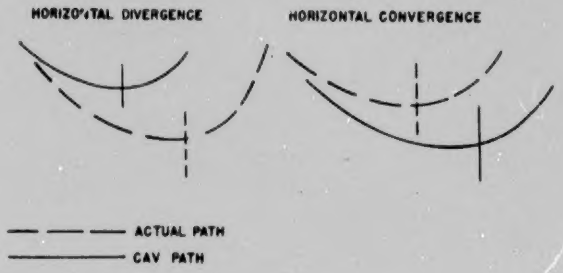

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Figure 4-39.—Modification of CAVT's because of convergence and divergence.

Evaluation of Vorticity

In addition to locating the areas of convergence and divergence in order to adjust the CAVT's we must also consider the effects of horizontal wind shear as it affects the relative vorticity, and hence the movement of the long waves and deepening or filling associated with this movement.

The two terms "curvature" and "shear," which determine the relative vorticity, may vary inversely to each other. Therefore, it is necessary to evaluate both of them. Figures 4-40 through 4-43 illustrate some of the possible combinations of curvature and shear. Solid lines are streamlines or contours; dashed lines are isotachs.

Figure 4-40 represents a symmetrical sinusoidal streamline pattern with isotachs parallel to contours. Therefore, there is no gradient of shear along the contours alone. In region I, the curvature becomes more anticyclonic downstream; therefore, relative vorticity decreases downstream and the whole region is favorable for deepening. The reverse is true west of the trough, region II, and the region is unfavorable for deepening.

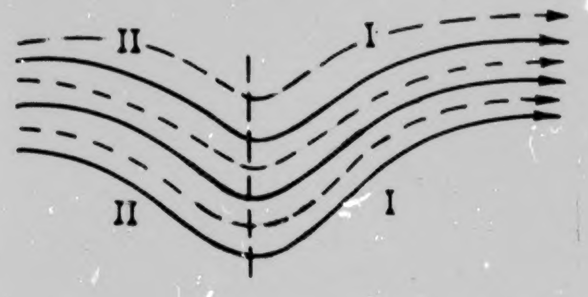

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Figure 4-40.–Contour-isotach pattern for shear analysis.

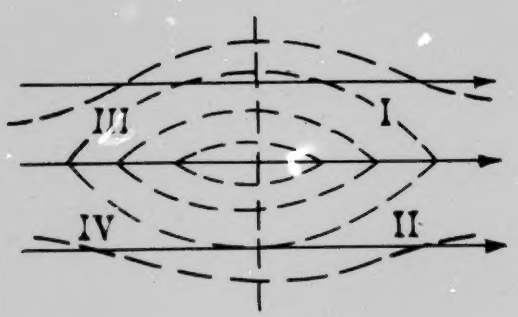

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Figure 4-41.–Contour-isotach pattern for shear analysis.

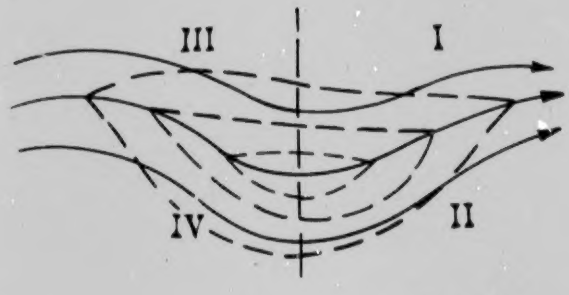

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Figure 4-42.–Contour-isotach pattern for shear analysis

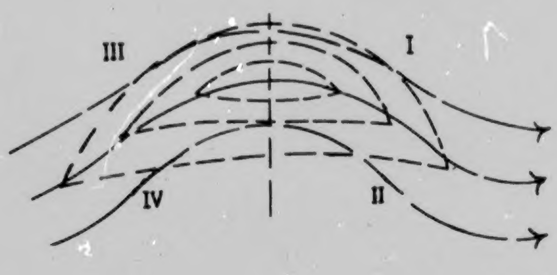

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Figure 4-43.–Contour-isotach pattern for shear analysis.

In figure 4-41 there is no curvature of streamlines; therefore, the shear alone determines the relative vorticity. The shear downstream in regions I and IV becomes less cyclonic; in regions II and III, it becomes more cyclonic. Regions I and IV are therefore favorable for deepening downstream.

In region I of figure 4-42 both cyclonic shear and curvature decrease downstream and this region is highly favorable for deepening. In region III both cyclonic shear and curvature increase downstream and this region is unfavorable for deepening. In region II the cyclonic curvature decreases downstream, but the cyclonic shear increases. This situation is indeterminate without calculation unless one term predominates. If the curvature gardient is large and the shear gradient small, the region is likely to be favorable for deepening. In region IV, the cyclonic curvature increases downstream, but the cyclonic shear decreases, so that this region is also indeterminate unless one of the two terms predominates.

In region I of figure 4-43 the cyclonic shear decreases downstream and the cyclonic curvature increases. The region is indeterminate; however, if the shear gradient is larger than the curvature gradient, deepening is favored. Region II has increasing cyclonic shear and curvature downstream and is quite unfavorable. In region III, the shear becomes more cyclonic downstream and the curvature becomes less cyclonic. This region is also indeterminate unless the curvature term predominates. In region IV, the shear and curvature become less cyclonic downstream and the region is favorable for deepening.

The whole point of these illustrations is to point out that decreasing cyclonic shear and curvature are favorable for deepening of the frontal wave and the reverse for increasing cyclonic shear and curvature and that when the situation is indeterminate, the perdominant term will determine whether or not there will be deepening. Deepening of a trough will increase the amplitude of the wave trough and shorten the wavelength, and the CAVT's must be adjusted accordingly. In fact, the shear analysis evaluation should be applied to the extrapolation of long wave movement and the movement by Petterssen's wave speed nomogram as well, for the reasons just stated. (This is discussed in chapter 8 of this training manual.)

Relation of Vorticity to Weather Processes

Vorticity not only affects the genesis of cyclones and anticyclones, but it also has a direct bearing on cloudiness, precipitation, and pressure and height changes as well. Vorticity is used primarily in forecasting cloudiness and precipitation over an extensive area. One rule states that when relative vorticity decreases downstream in the upper troposphere, convergence is taking place in the lower levels. When convergence takes place, cloudiness and possibly precipitation will prevail if sufficient moisture is present.

One rule using vorticity in relation to cyclone development stems from the observation that when cyclone development occurs, the location almost without variance is in advance of an upper trough. Thus, when an upper level trough with positive vorticity advection in front of it overtakes a frontal system in the lower troposphere, there is a reliable indicator of cyclone development on the surface chart. This is usually accompanied by deepening and intensification of the surface system. Also, the development of cyclones at sea level takes place when and where an area of appreciable positive vorticity advection situated in the upper troposphere overlies a slow moving or quasi-stationary front on the surface chart.

It has been shown before that surface pressure changes that take place aloft. The relationship between convergence and divergence can best be illustrated by the shear term. If we consider a flow where the cyclonic shear is decreasing downstream (stronger wind to the right than to the left of the current), more air is being removed from the area than is being fed into it, hence a net depletion of mass aloft, or divergence. Divergence aloft is associated with surface pressure falls, and since this is the situation, the relative vorticity is decreasing downstream. We may state that surface pressure falls where relative vorticity decreases downstream in the upper troposphere, or where advection of more cyclonic vorticity takes place aloft. The converse of this is in the case of convergence aloft.

CHAPTER 5

AIR MASSES, FRONTS, AND CYCLONES

In the preceding chapters you have seen that regions of the earth have special properties of temperature and humidity. Under favorable conditions these regions impart these properties to the overlying air, thereby forming air masses. These air masses have different properties depending upon whether they form over land or water surfaces or over polar or tropical regions. When two air masses with different properties are brought together a line of discontinuity or front is formed between them. When a warm air mass lies adjacent to a cold air mass and either of these is caused to be accelerated along a part of the front, there is a tendency for a wave motion to be set up along the front. The result of this interaction is called a wave cyclone. The wave cyclone may be either stable, that is, it travels along the front in a horizontal plane without appreciably intensifying; or it may be unstable, thereby going through a life cycle—occluding and dying.

The material contained in this chapter is an extension to that of chapter 6, Air Masses and Fronts, Aerographer's Mate 3 & 2, NavTra 10363-D. A review of the basic material would be helpful in understanding some of the more advanced material in this chapter.

AIR MASSES

CONDITIONS NECESSARY FOR FORMATION

Two primary factors for the production of air masses are as follows: First, a surface whose properties, essentially temperature and moisture, are relatively uniform (it may be a water, land, or snow covered area); and second, a large divergent flow which tends to destroy temperature contrasts and produces a homogeneous mass of air. The energy supplied to the earth's surface from the sun is distributed to the air mass by convection, radiation, and conduction.

Another necessary condition for air mass formation is equilibrium between ground and air. This is established by a combination of the following processes: (1) turbulent-convective transport of heat upward into the higher levels of the air; (2) cooling of air by radiation loss of heat; and (3) transport of heat by evaporation and condensation processes.

By far the most effective process involved in establishing equilibrium is turbulent-convective transport of heat upwards. It is also the fastest process whereby equilibrium is established. The least effective and slowest process whereby equilibrium is established is radiation.

Evaporation and condensation serve to contribute greatly, during both radiation and turbulent-convective processes, in conserving the heat of the overlying air by the water vapor permitting radiation only through transparent bands in the case of radiation cooling and in releasing the latent heat of condensation in the case of turbulent-convective processes. For this reason, Aerographer's Mates can readily see that the tropical latitudes form air masses rapidly primarily through the upward transport of heat by the turbulent-convective process and the polar regions form air masses slowly primarily by the loss of heat through radiation.

The Aerographer's Mate will further recognize that the underlying surface is not uniform in its thermal properties throughout the year and the distribution of land and water is unequal, and that for these reasons special summer or winter air masses may be formed. The rate of air mass formation will further vary with the intensity of insolation.

Effects of Air Mass Circulation

It is recognized that there are three circulation types over the earth. Some of these are favorable for air mass development, while others do not fulfill the requirements. They are:

1. The anticyclonic systems of the earth. They have stagnant or slowly moving air which allows time for air to adjust its heat and moisture content to that of the underlying surface. These anticyclones possess a divergent airflow which spreads the properties over a large area horizontally, with turbulence and convection distributing these properties vertically. Subsidence, another property of anticyclones, is favorable for lateral mixing which results in horizontal or layer homogeneity.

   Warm highs extend to great heights due to a lesser density gradient aloft and thereby produce an air mass of relatively great vertical extent. Two examples of warm highs are the Bermuda high and the Pacific high. Cold highs are of moderate or shallow vertical extent and thereby produce air masses of moderate or shallow height.

2. Cyclonic systems are not conducive to air mass formation because they are characterized by greater wind speeds than anticyclonic systems. These wind speeds prevent cyclonic systems from stabilizing the necessary amount of time an air mass should remain over a surface to adjust itself. An exception is the heat low

3. Belts of convergence are not conducive to air mass formation since they have essentially the same properties as cyclonic systems.

There are two areas of convergence which are exceptions to this rule. These are the area over the north Pacific between Siberia and North America and the area over the Atlantic off the coast of Labrador and Newfoundland. These two areas act as source regions for maritime polar air.

SOURCE REGIONS

Source regions are areas of the earth's surface which possess the necessary conditions conducive to air mass formation. These conditions were described in the preceding paragraphs.

Generally speaking, there are two primary air mass source regions. One is associated with the stagnant cold polar regions where the cold air masses develop; and the other exists along the lower latitudes near 30° where the warm air masses develop. Between these two areas there exists a zone of transition over which mixing of the cold air masses from the north and warm air masses from the south takes place.

A more detailed description of air mass source regions may be found in AG 3 & 2, chapter 6.

CLASSIFICATION OF AIR MASSES

The source region has been found to be a useful criterion as a first consideration when classifying air masses. The letter designations used to indicate the various source regions are as follows:

A – Arctic/Antarctic air masses.
P – Polar air masses.
T – Tropical air masses.
E – Equatorial air masses.
S – Superior air masses.

The second criterion is made by prefixing an "m" or "c" to one of the above to indicate whether the air mass has maritime or continental characteristics.

It should be noticed that, due to the predominance of land masses or ice fields in the Arctic, maritime Arctic (mA) would be a rare type. In a like manner, equatorial air (E) is found exclusively over the ocean surface in the vicinity of the Equator and is designated neither "c" nor "m" but simply E.

A third letter is added to indicate whether the air is colder (k) or warmer (w) than the surface over which it is moving. This is the thermodynamic classification. The "k" type air mass will be warmed from below, and convection currents will form. Some characteristics of this type of air mass are turbulence up to about 10,000 feet; unstable lapse rate (nearly dry adiabatic); good visibility except in showers and dust storms; cumulonimbus clouds such as cumulus and cumulonimbus; and showers, thunderstorms, hail, ice pellets, and snow flurries.

The "w" type of air mass is cooled from below. This type tends to maintain its original properties and is modified only in the lower few thousand feet as it moves. Some characteristics of this type are smooth air (above the friction level); stable lapse rate; poor visibility (smoke and soot are held in the lower few thousand feet); and stratiform clouds and fog and occasionally drizzle.

Properties of Air Masses in Their Source Regions

The properties an air mass will acquire in its source region are dependent upon a number of factors–the time of year (winter or summer), the nature of the underlying surface (whether land, water, or ice covered), and the length of time it remains over its source region. The properties described in the following section are, in general, applicable to all areas of the world, except when the differences are pointed out. You must remember that these indicate only average or normal properties and conditions, and those characteristics and properties may be different on any particular day.

Arctic and Polar Continental Air Masses

Since these two air masses have similar characteristics, they are discussed together. In the United States continental polar air (cP) is usually a modification of cA, although it may be more unstable in the eastern part of the United States. In addition, cP air is the air mass that usually interacts with mT air along the polar front, and resultant rains and snows make up a large proportion of the winter precipitation from the Rocky Mountains eastward.

In the winter, its North American location is from the Atlantic Ocean to the Bering Sea from 50° north to the Arctic Circle. Over Siberia, it is from the Pacific Ocean to the edge of Europe and the Arctic Ocean south to the Plateau of Tibet. Over Europe, a more distinct separation between Arctic and continental polar air is due to the cold easterly currents toward the Icelandic low from the Arctic regions and the advection of warmer air from the Atlantic Ocean easterly toward the Barents Sea.

The lower layer of continental polar air in its source region is characterized by a strong surface inversion, an isothermal middle layer, and a steep lapse rate aloft.

MOISTURE DISTRIBUTION.–Moisture adapts to temperature so as not to exceed saturation. Moisture in the lower layers is withdrawn by frost and ice crystal fogs.

According to Wexler, all polar air masses above 10,000 to 15,000 feet are alike. The essential difference between continental polar and maritime polar is the modification up to that level, and modification is dependent upon the snow or water source regions. Above this level, a steep lapse rate is exhibited in both air masses. This air aloft is sometimes referred to as polar superior air.

The mixing ratio is low, usually less than 1 gram per kilogram. The humidity is high at the surface (85-95 percent), decreasing to the 50's in the isothermal layer and lower aloft. The effect of cooling, however, seldom exceeds 10,000 feet in the cA air masses that reach the United States.

In general, the skies will be clear or have flattened stratocumulus clouds, if any. Visibility will be good, except if ice crystal fogs are present (in temperatures −30°C and below). Refer to figure 5-1 for typical lapse rates of cP air in its source region.

Continental polar air has the greatest annual variation of any air mass, and its properties show a large variation from season to season. The warming may result in a weaker inversion in summer that may completely disappear during the day only to be reestablished at night due to cooling. European continental polar air is more moist than North American or Siberian continental air. Over Asia, cP air is colder than other cP air masses.

Occasionally we have intrusions of maritime polar air into the continental polar source regions. A low in the Beaufort Sea region will induce maritime polar air into the Mackenzie River Valley by bringing it from the Bering Sea across northern Alaska. Temperatures will rise from 20° to 40°F, but this air is quickly modified to continental polar air. Another case is that of the Nova Scotia wave. A strong low in the vicinity of Nova Scotia will induce mP from the North Atlantic into the continent in the

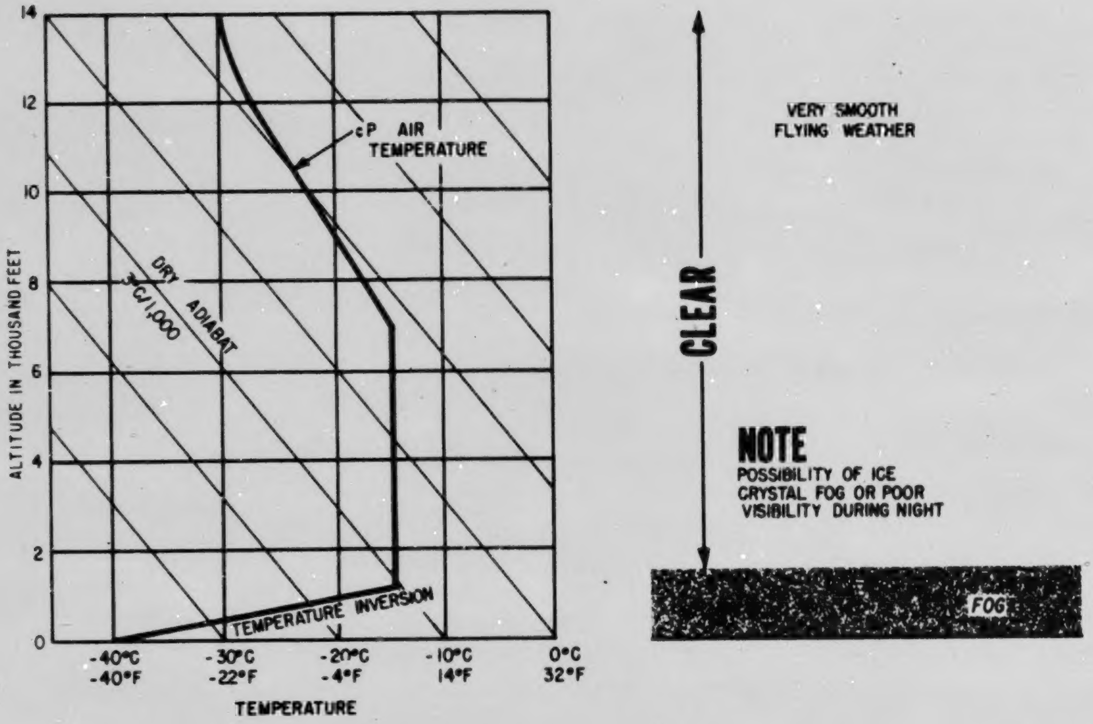

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Figure 5-1.—Continental polar air in its source region (winter).

Labrador-Hudson Bay region. Temperatures in this region will rise 30° to 40°F and the air will be modified quickly to cP, though not as rapidly as air in the Mackenzie Valley.

In summer, due to extremely long days and the absence of a snow cover over a large portion of the source region, cP air masses are very different from those observed during the winter months. The warmth of the underlying surface produces unstable layers near the ground in contrast to the extremely stable conditions during winter.

When an air mass originating over the Arctic Ocean in summer moves southward, it is impossible to distinguish it after it reaches the United States from one originating over continental North America. During this season, even mP air from the North Pacific Ocean becomes practically identical with the cP masses after crossing the western mountains of the United States. Continental polar air masses in summer are characterized by relatively low temperatures and are subject to wide daily variations due to their characteristic dryness aloft. During afternoon hours, adiabatic or superadiabatic lapse rates may develop near the ground and produce turbulence as high as 6,000 to 10,000 feet. Occasionally, after the air has stagnated for a few days over the southeastern United States, sufficient moisture may accumulate to result in the formation of local thundershowers, usually confined to the mountainous districts. Over Europe, the source region is essentially the same as for the winter months, but it is predominantly a dry air mass and generally produces fair weather, except for haze and smoke present in the lower layers.

Maritime Polar Air

Maritime polar air masses are formed in transition zones, and the properties the air mass will have are dependent upon its original source and its trajectory through the transition zone.

The air may arrive at the transition zone either by a path from over land to the water, which may be either cyclonic or anticyclonic, or from an overwater path from the subtropical regions.

If the air arrives at the source region in winter from a continental source over the transition zone and the trajectory is cyclonic, the air is warmed from below, heat is added by radiation and eddy exchange from the surface, and moisture is added which is carried aloft by convection. Subsidence is not operative as the air is usually dominated by the low which is associated with the outbreak. Modification usually extends up to about 12,000 feet. When the air arrives over the eastern portion of the mP source region, its stability increases because the rate of heating from below is reduced. The air is usually dry and cold aloft with a continental character.

On the other hand, if the air moves out from its continental source and passes through the transition zones with an anticyclonic trajectory, the air is more stable, particularly at upper levels, due to the subsidence effect.

Also, in winter, air may arrive at the mP source region from the subtropics due to an elongated trough extending southward; the air has a long overwater trajectory and is cooled by its passage over colder water on its path northward. In this case, the air is cooled in its lower layers and therefore more stable. This is the warmest air that arrives in this region.

In general, maritime polar air in its source region in the winter is convectively unstable in its lower layers, dry and cold aloft and has little or no diurnal temperature variation. Mild convective type clouds and precipitation predominate.

Just as there is an appreciable difference in the structure of cP air masses in summer as compared with winter, so we find pronounced differences in the structure of polar maritime air. In summer, the maritime air masses are more stable than winter.

In summer, air arriving at its mP source region from continental regions is cooled from below, resulting in extreme stability in the lower layers. Fogs and stratus are common.

Maritime Tropical

The subtropical highs are the source regions for mT air. On the eastern side, colder temperatures and marked subsidence render this portion of the high stable. The western portion is the most unstable due to warmer surface conditions and a lesser effect of subsidence. Most often, mT air is modified mP air, but all air masses are continually acquiring new properties in primary and secondary source regions. Subsidence, especially over the eastern limbs of the high, prevents the distribution of moisture to high levels. (See figs. 2-6 and 2-7 in chapter 2 of this manual.) Therefore, the air is moist below the inversion and dry above. (See Trade Inversions in chapter 12.) Cloudiness is more prevalent in the western portion, and the amount of precipitation will be small unless associated with frontal convergence or some other type of convergence. Typical properties of mT air over the Atlantic in winter in its source region are a moderately steep lapse rate with the air convectively unstable to about 9,000 to 11,000 feet with a mixing ratio of 13 to 15 grams per kilogram. Relative humidity is high in the lower levels and decreases aloft. Surface temperatures average around 75° to 80°F and the 800-mb temperature about 12° to 13°C. Therefore, the air is characterized in its source region by high temperature, high moisture content, conditional instability, and a few stratocumulus clouds.

Summer conditions are very similar to those of winter except that the increased surface temperature in the western portion causes increased instability and air mass showers and thunderstorms. In the eastern portion, the axis and latitude of the seasonal shift of the subtropical high-pressure cells cause a flow which results in upwelling along the western coasts of continents, with increased stability and the prevalance of coastal stratus and fogs.

Continental Tropical

In winter this air mass is characterized by moderately warm temperatures, but in summer they are extremely warm. This is the warmest air mass found in North America. Due to the usual extension of a high-pressure cell over these regions in the winter, the flow is anticyclonic and the air consequently stable. In summer the dry land is strongly heated. There is usually a thermal low present at the surface with a high level anticyclone aloft. This air mass is characterized by strong daytime convection in the summer with a strong and steep lapse rate, often dry adiabatic to high levels.

During winter and summer the air has a high condensation level, due to the lack of moisture, and clouds are rare. A few small cumulus may form late in the afternoon. The air, especially in summer, has a large diurnal temperature variation, often as high as 30° to 40°F.

Equatorial Air

There is little or no seasonal variation in the properties of this air mass. It is characterized by a great uniformity in the distribution of temperature and humidity. The temperatures are high and instability is the rule. Humidity is also high and uniformly distributed, resulting in a low condensation level. When occurring over land or when forced aloft by some kind of lifting action, the weather is characterized by convective instability, producing showers or thunderstorms.

Superior Air

Superior air is a high level air mass and has little direct relation to the air mass which it is overlying. It develops above an inversion layer. Superior air is created by an anticyclonic flow and subsidence, and is very stable and dry. It is warmer than any other air mass for its altitude, with little seasonal variation in its properties.

AIR MASS MODIFICATION

When an air mass moves out of its source region, there are a number of factors which act upon the air mass to change its properties. These modifying influences do not occur separately. For instance, in the passage of cold air over warmer water surfaces, there is not only a release of heat to the air, but also some moisture. The Aerographer's Mate must be aware of the changes that take place once an air mass has left its source region in order to integrate these changes into his forecast.

MODIFYING FACTORS

As an air mass expands and slowly moves out of its source region, it travels along a certain path. The surface over which this path takes the air mass after leaving its source modifies the air mass. The type trajectory, whether cyclonic or anticyclonic, also has a bearing on its modification. The time the air mass has been out of its source region will determine to a great extent the characteristics of the air mass when a thermodynamic classification is attempted.

Surface Conditions

The first modifying factor on an air mass as it leaves its source region is the type and condition of the surface over which it travels. Here, the factors of surface temperature, moisture, and topography must be considered.

TEMPERATURE.—The temperature of the surface relative to that of the air mass will modify not only the temperature, but its stability as well. For example, if the air mass is warm and moves over a colder surface, such as tropical air moving over colder water, the cold surface cools the lower layers of the air mass and its stability is increased. This stability will extend to the upper layers in time, and condensation in the form of fog or low stratus normally occurs. (See fig. 5-2.)

If the air mass moves over a surface that is warmer, such as polar continental air moving out from the continent in winter over warmer water, the warm water heats the lower layers of the air mass, increasing instability and consequently spreading to higher layers. Figure 5-3 illustrates the movement of cP air over a warmer water surface in winter.

The changes in stability of the air mass give valuable indications of the cloud types that will form, as well as the type of precipitation. Also, the increase or decrease in stability gives further indications of the lower layer turbulence and visibility.

MOISTURE.—The air mass may be modified in its moisture content by the addition of moisture by evaporation or by the removal of moisture by condensation and precipitation. If the air mass is moving over continental regions, the existence of unfrozen bodies of water can greatly modify the air mass and, in the case of an air mass moving from a continent to an ocean, the modification can be considerable. In general, dependent upon the temperature of the

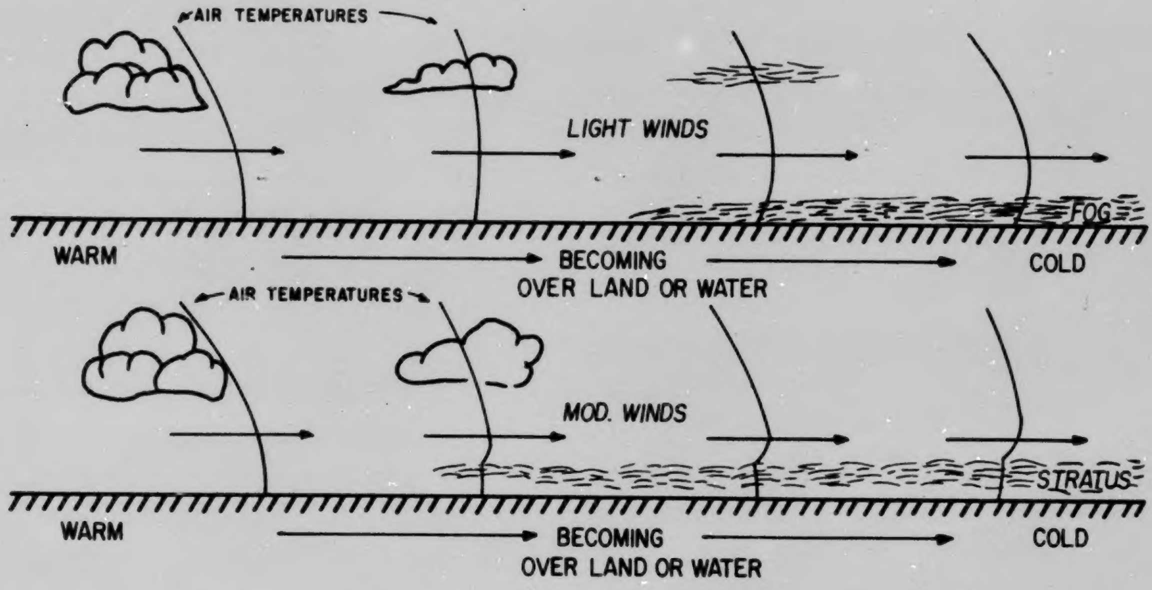
AIR TEMPERATURES
LIGHT WINDS
FOG
WARM
BECOMING
COLD
OVER LAND OR WATER
AIR TEMPERATURES
MOD. WINDS
STRATUS
WARM
BECOMING
COLD
OVER LAND OR WATER

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Figure 5-2.—Passage of warm air over colder surfaces.

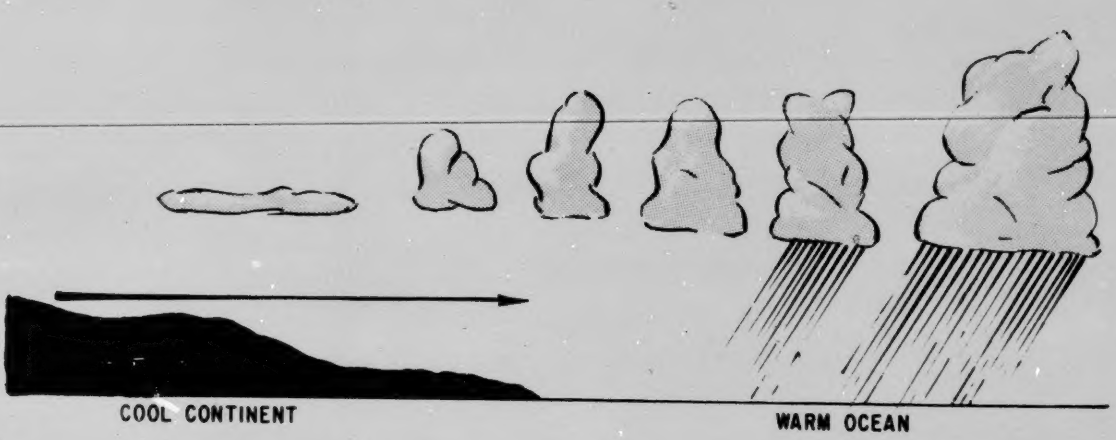
COOL CONTINENT
WARM OCEAN

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Figure 5-3.—Continental polar air moving from cool continent to warm ocean (winter).

two surfaces, the movement over a water surface will increase the moisture content of the lower layers, and the relative temperature of the surface.

For example, the passage of cold air over a warm water surface will decrease the stability of the air with resultant vertical currents. The passage of warm moist air over a cold surface increases the stability and could result in fog as the air is cooled and mositure is added by evaporation.

TOPOGRAPHY.–The effect of topography is evident primarily in the regions of mountains. The air mass is modified on the windward side by the removal of moisture through precipitation with a decrease in stability; and as the air descends on the other side of the mountain, the stability increases as the air becomes warmer and drier.

Trajectory

After an air mass has left its source region, the trajectory it follows, whether cyclonic or anticyclonic, has a great effect on its stability. If the air follows a cyclonic trajectory, its stability in the upper levels is decreased; this instability is a reflection of the cyclonic relative vorticity. The stability of the lower layers is not greatly affected by this process. On the other hand, if the trajectory is anticyclonic, its stability in the upper levels is increased as a result of subsidence associated with anticyclonic relative vorticity.

Age

Although the age of an air mass in itself cannot modify the air mass, it will determine, to a great extent, the amount of modification that takes place. For example, an air mass that has recently moved from its source region will not have had time to become modified significantly. However, an air mass which has moved into a new region and stagnated for some time, and is now old, will be found to have lost many of its original characteristics.

Summary

In figure 5-4 the two modifying influences are classified thermal and mechanical.

THE PROCESS	HOW IT HAPPENS	RESULTS
A. THERMAL		
1. Heating from below.	Air mass passes from over a cold surface to a warm surface, or surface under air mass is heated by sun.	Decrease in stability.
2. Cooling from below.	Air mass passes from over a warm surface to a cold surface, OR radiational cooling of surface under air mass takes place.	Increase in stability.
3. Addition of moisture.	By evaporation from water, ice, or snow surfaces, or moist ground, or from raindrops or other precipitation which falls from overrunning saturated air currents.	Decrease in stability.
4. Removal of moisture.	By condensation and precipitation from the air mass.	Increase in stability.
B. MECHANICAL		
1. Turbulent mixing.	Up- and down-draft.	Tends to result in a thorough mixing of the air through the layer where the turbulence exists.
2. Sinking.	Movement down from above colder air masses or descent from high elevations to lowlands, subsidence and lateral spreading.	Increases stability.
3. Lifting.	Movement up over colder air masses or over elevations of land or to compensate for air at the same level converging.	Decreases stability.

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Figure 5-4.—Air mass changes.

The figure indicates the modifying process, what takes place, and the resultant change in stability of the air mass. It must be reiterated that these processes do not occur independently, but two or more processes are usually in evidence at the same time.

It should be pointed out that within any single air mass the weather is controlled by the moisture content, stability, and the up-and-down slope movements of air. It must also be stressed that the conditions indicated are only average conditions and that each individual case may be quite different.

CHARACTERISTICS OF MODIFIED AIR MASSES

North American Air Masses (Winter Season)

CONTINENTAL ARCTIC (cA) AND CONTINENTAL POLAR (cP).—Figure 5-5 illustrates some of the paths taken by cA and cP air entering the United States.

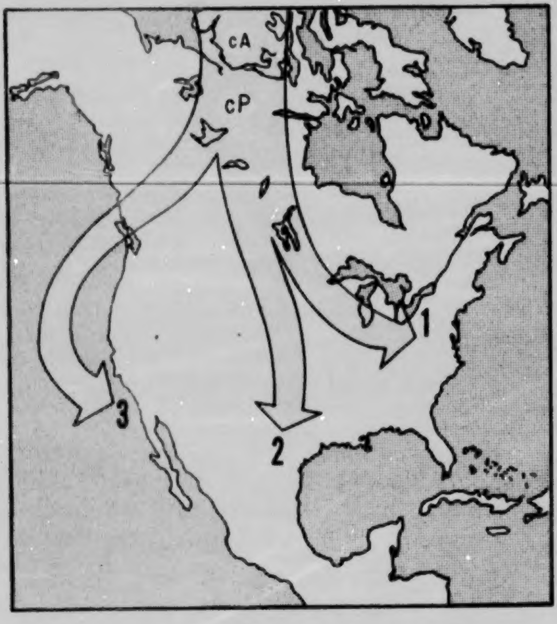

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Figure 5-5.—Paths of cP and cA air (North American winter).

Path No. 1 (cyclonic) is usually indicative of a strong outbreak of cold air accompanied in the lower few thousand feet by high winds and turbulence with gusty flying conditions. A strong inversion persists between 5,000 and 10,000 feet, and fractocumulus and stratocumulus clouds often form under this inversion. As the air mass, following this path, moves over the Great Lakes, it is heated from below and moisture is added, especially during early winter before the lakes have frozen over. This results in either rain or snow showers on the lee side of the Great Lakes and on the windward side of the Appalachians. (See fig. 5-6.)

East of the mountains, relatively clear skies prevail. The cloud bases are at 500 to 1,000 feet on the lee side of the Great Lakes and at or near zero over the mountains. Cloud tops are at about 7,000 feet in the Great Lakes region and around 14,000 feet in the mountain region. (See fig. 5-6.)

In the Middle West, clouds associated with this type of air mass continue for 24 to 48 hours after the arrival of the cold mass, while along the Atlantic Coast rapid passage of the leading edge of a cA air mass produces almost immediate clearing.

When polar air follows path No. 2 (anticyclonic) over the central United States, generally smooth flying conditions exist, except in the lower 4,000 feet where contact of the air with the warmer surface may cause turbulence. Surface visibility is good, except during the early morning hours when the air mass stagnates over a region and subsidence occurs aloft.

Path No. 3 does not occur very often. This is essentially a wintertime phenomenon and usually occurs when an intense high in the cP source region (1,035 to 1,040 mb) results in a strong pressure gradient forcing air over the mountains. When the trajectory is similar to path No. 3, the air arrives over the central and southern coast of California as a cold, convectively unstable current with subsequent squalls, rain showers, and even snow occurring as far south as southern California.

MARITIME ARCTIC (mA) AND MARITIME POLAR (mP) (PACIFIC).—When an outbreak of polar air moves over only a very small part of the Pacific Ocean before reaching the United States, it usually resembles mAk. If its path has

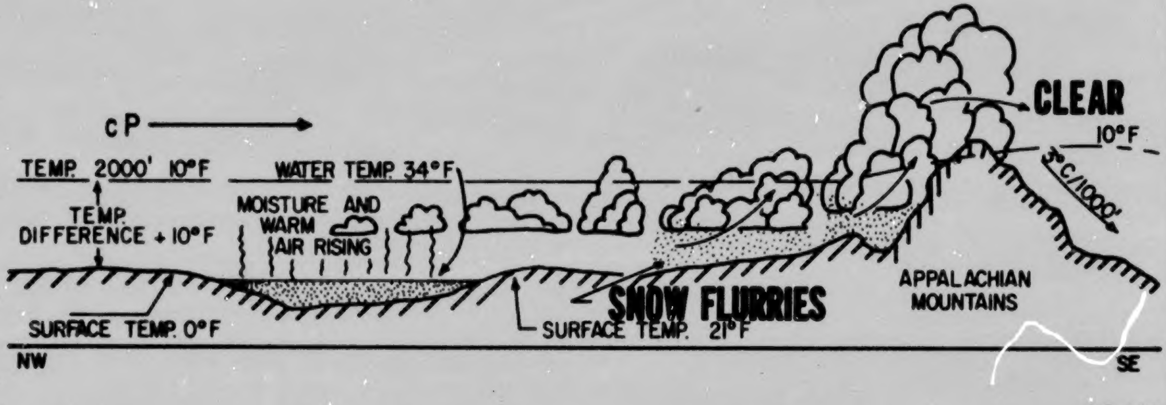
cP
TEMP. 2000' 10°F
TEMP. DIFFERENCE + 10°F
SURFACE TEMP. 0°F
WATER TEMP. 34°F
MOISTURE AND WARM AIR RISING
SNOW FLURRIES
SURFACE TEMP. 21°F
APPALACHIAN MOUNTAINS
CLEAR
10°F
3°C/1000'
NW
SE
AG.460

Figure 5-6.—cP air moving over the Great Lakes (winter).

been far to the south, it is typically mP. Figure 5-7 shows some of the trajectories by which mP air reaches the North American coast during the winter.

Path No. 1 is a cyclonic trajectory. The air originates in Alaska or northern Canada and is pulled out over the Pacific Ocean by a low close to British Columbia in the Gulf of Alaska. This air has a relatively short overwater path and brings very cold weather to the Pacific Northwest. The air is convectively unstable in the lower layers; and when the air is lifted by the coastal ranges, showers and squalls are common with 2,000-foot ceilings along the coast and near zero along the mountain ranges. The visibility is good except in precipitation. Rough flying weather prevails in the turbulent air. Icing is most noticeable in the mountains and may be severe.

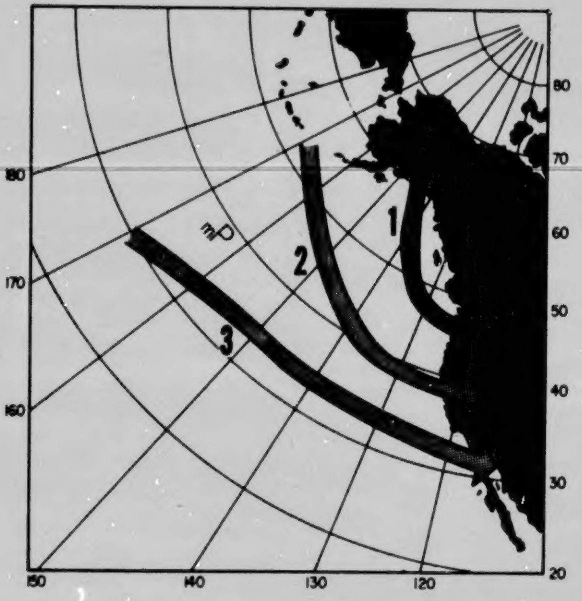
80
70
60
50
40
30
20
180
170
160
150
140
130
120
mP
1
2
3
AG.461

Figure 5-7.—Paths of mP air over the Pacific coast in winter.

Maritime polar (mP) air with a longer overwater trajectory path No. 2 dominates the west coast of the United States during winter months. When there is rapid west to east motion and small north to south motion of pressure systems, mP air may influence the weather over most of the United States. Due to a longer overwater trajectory, this mP air is heated to greater heights, and convective instability is present up to about 10,000 feet. This air has typical "k" characteristics-turbulent gusty winds, steep lapse rate, good visibility at ground except in precipitation, and cumulus and cumulonimbus clouds with showers. These showers are not as intense as those produced in the shorter trajectory mP air, but the total amount of precipitation is greater. With a very long overwater trajectory, path No. 3, mP air tends to become rather stable, with one or two subsidence inversions and stratus type clouds with base about 1,000 feet and tops about 4,000 feet.

Maritime polar (mP) air sometimes stagnates between the mountains in the Great Basin region of the Western United States. This cold air becomes modified, and subsidence inversions, low stratus clouds, and fog form. This process causes the Pacific coast valleys to be among the foggiest places in the country during winter. When this air moves eastward across the Continental Divide and downslope, it generally brings clear skies, mild temperatures, and low humidities.

Maritime polar air that moves eastward without stagnating has much of its moisture condensed out during the lifting necessary to cross the mountains. It then warms dry adiabatically as it descends, forming a very stable air mass with good flying conditions. This air has large daily temperature ranges. When mP air crosses the Rocky Mountains and encounters a deep cP air mass which forces it aloft, very severe snowstorms or blizzards may occur.

MARITIME POLAR (ATLANTIC ORIGIN).–The rarest air mass in the United States is mP air of Atlantic origin, owing to the normal westeast movement of systems. The air mass is confined to the east coast of the United States, New England, and the Maritime Provinces of Canada. The air mass is originally cP which has had a short trajectory over water, but the heating, addition of moisture, and steepening of lapse rate is not as great as in the mP air of Pacific origin.

MARITIME TROPICAL (PACIFIC ORIGIN).–Maritime tropical air of Pacific origin is seldom observed along the west coast of the United States, especially near the ground. On those occasions when true mT air moves over the lower coast of California, the air is pushed aloft in the rapidly occluding frontal systems, producing the heaviest winter rains on record there.

MARITIME TROPICAL (ATLANTIC ORIGIN).–The temperature and moisture content are higher than in any other air mass on the North American Continent in winter. Maritime tropical air is responsible for the greatest portion of precipitation both in summer and in winter. Most of the precipitation falls through a polar air mass which it is overrunning and is thus associated with frontal activity.

In the Southern States, mT air is mild with much low cloudiness at night and a high frequency of fogs. During the day convective activity raises ceilings; flying conditions are good. As mT air moves northward, extensive cooling in the lower layers takes place, causing much fog and low stratus, which sometimes precipitates a fine drizzle. True mT air in winter has a dewpoint in excess of 60°F and should not be confused with modified cP returning from the Gulf of Mexico. Northward movement of mT occurs only with a very low zonal index situation.

Summer Air Masses

Most of the United States is dominated by either S or mT air, whereas Canada (and Northwestern United States) is dominated by polar air. Occasionally, tropical air is transported to the Canadian tundra and Hudson Bay region.

MARITIME POLAR (PACIFIC ORIGIN).–The whole of the Pacific coast is usually under the influence of mP air in the summer. This air seldom extends east of the Rockies; and when it does, it has acquired properties almost identical to cP air. Coastal weather is generally clear with scattered cumulus. Typical unstable conditions prevail along the Pacific coast as far as northern California with the influx of such mP air.

MARITIME POLAR (ATLANTIC ORIGIN).–The east coast of the United States occasionally experiences an influx of mP air from the Atlantic, but due to the lesser temperature contrast between continent and ocean in summer, this influx of mP air creates no hazard to flying. The temperature drops quite a bit, bringing relief from heat waves. Precipitation from this air is rare.

CONTINENTAL POLAR.–Polar air of Canadian origin occasionally invades the United States in summer under a low zonal index situation. When it does, it preserves to a large extent its temperature characteristics. Convective activity in the United States is extensive but mild, confined in height to 700 mb or less.

MARITIME TROPICAL (ATLANTIC ORIGIN).–The most extensive air mass in summer over the Eastern United States. High temperature and a great deal of moisture characterize this air, with a summer dewpoint near or in excess of 70°F at the surface. Movement of mT air over warmer ground increases its convective instability. Low stratiform clouds are the rule in the mornings, especially along the east coast, becoming convective clouds during the day, with frequent thunderstorms at night. Flying conditions in this air mass are not hazardous despite the thunderstorms because they are easily circumnavigated. Ground fogs are frequent with northward movement of mT air over land, and sea fogs with movement over water. The famous fogs of the Grand Banks are typical of mT air over a cold ocean current.

Occasionally during late summer and early fall, the Bermuda high forms a westward extension going as far west as Lower California. The intense surface heating and orographic lifting makes this situation intensely unstable with thunderstorms of cloudburst intensity (SONORA weather).

CONTINENTAL TROPICAL.—This air is found only during the summer forming over a small area of northern Mexico, western Texas, New Mexico, and eastern Arizona. It can be identified by its extremely high surface temperatures, very low humidities, large diurnal-temperature ranges, and infrequent precipitation. Flying conditions are excellent with respect to weather, but clear air turbulence makes for a rough ride.

Figure 5-8(A) and (B) show the properties of North American air from the standpoint of flying during the winter and summer seasons.

Air Masses of Europe

The air masses of Europe do not fit neatly into the same categories as those of North America. Prinicipally, this is due to the fact that the atmospheric separations occur at a different range of values of temperature. This is because the geographical environments of the two continents differ greatly. The continent of North America widens toward the north; the opposite is true of Europe. The great mountain systems of North America run north and south; those of Europe are oriented east and west.

CONTINENTAL ARCTIC.—Continental Arctic (cA) air comes into Europe from the northeast. It is much more frequently felt in eastern Europe. In mild winters it affects western Europe very little. Severe winters in western Europe are due to cA air being abnormally displaced to the west. In the warmer months it occurs very infrequently in western Europe.

The cA air in Europe undergoes a great deal of modification as it spreads out over the continent, especially in the west, where the air, even in winter, shows the results of a great deal of mixing and stirring. However, in and near the source region there is a great deal of stability in the low layers.

MARITIME ARCTIC.—Maritime Arctic (mA) air masses spread into Europe from the north. Due to its rapid movement over warm seas, it becomes quite unstable (especially in winter) with a lapse rate close to the saturated adiabatic rate in all layers. The air is not greatly different from maritime polar air, whose source region is to the west. It has "k" characteristics only. Outbreaks of this air (mAk) occur mostly in winter, with only occasional outbreaks of such air in the summer, and then it is quite shallow and undergoes modification quickly.

In winter, cold rain or snow showers accompany fresh outbreaks of mA air, along with a great deal of turbulence. Where orographic lifting takes place, the snowfall is extremely heavy. These mA air masses are sometimes the cause of the MISTRAL in southern France.

MARITIME POLAR.—Maritime polar (mP) air masses predominate in western Europe. They are quite varied, dependent on their source regions and dynamic effects. With cyclonic circulation they show noticeably the effects of stirring and convergence and a steep lapse rate; with anticyclonic circulation subsidence and divergence tend to cause the air mass to have a more stable lapse rate.

Mild showers accompany fresh outbreaks of mP air with "k" characteristics. However, as the air spreads over the continent, it becomes stable in winter. It becomes more unstable in the summer; but due to continental heating, it becomes too dry to cause appreciable shower activity.

(A)

AIR MASS	CLOUDS	CEILINGS	VISIBILITIES	TURBULENCE	SURFACE TEMPERATURE F.
cP(near source region)	None	Unlimited	Excellent(except near industrial areas, then 1 - 4 miles).	Smooth except with high winds velocities.	-10 to -60.
cP(southeast of Great Lakes)	Stratocumulus and cumulus tops 7,000-10,000 feet.	500-1,000 feet, 0 over mountains.	1 - 5 miles, 0 in snow flurries.	Moderate turbulence up to 10,000 feet.	0 to 20.
mP(on Pacific coast)	Cumulus tops above 20,000 feet	1,000-3,000 feet, 0 over mountains.	Good except 0 over mountains and in showers.	Moderate to strong turbulence.	45 to 55.
mP(east of Rockies)	None	Unlimited	Excellent except near industrial areas, then 1 - 4 miles.	Smooth except in lower levels with high winds.	30 to 40.
mP(east coast)	Stratocumulus and stratus tops 6,000-8,000 feet.	0-1,000 feet	Fair except 0 in precipitation area.	Rough in lower levels.	30 to 40.
mT(Pacific coast)	Stratus or stratocumulus.	500-1,500 feet	Good	Smooth	55 to 60.
mT(east of Rockies)	Stratus or stratocumulus.	100-1,5000 feet	Good	Smooth	60 to 70

(B)

AIR MASS	CLOUDS	CEILINGS	VISIBILITIES	TURBULENCE	SURFACE TEMPERATURE F.
cP(near source region)	Scattered cumulus	Unlimited	Good	Moderate turbulence up to 10,000 feet.	55-60
mP(Pacific coast)	Stratus tops, 2,000-5,000 feet	100 feet-2,500 feet, unlimited during day	1/2 - 10 miles	Slightly rough in clouds. Smooth above.	50-60
mP(east of Pacific)	None except scattered cumulus near mountains	Unlimited	Excellent	Generally smooth except over desert regions in afternoon.	60-70
S(Mississippi Valley)	None	Unlimited	Excellent	Slightly rough up to 15,000 feet.	75-85
mT(east of Rockies)	Stratocumulus early morning; cumulonimbus afternoon.	500-1,00 feet a.m.; 3,000-4,000 feet p.m.	Excellent	Smooth except in thunderstorms, then strong turbulence.	75-85

Figure 5-8.—Properties of air masses over North America from the standpoint of flying, (A) Winter; (B) Summer.

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Occasionally mP air spreads across France and into the Mediterranean Sea between the Pyreness and the Alps and travels eastward as far as the northwestern coast of India without losing its identity.

CONTINENTAL POLAR.–The continental polar (cP) air masses of Europe are sort of neutral air masses between the Arctic air and the tropical air. Quite frequently they are simply modification of cA air. The cP air is especially widespread when a retrograde or stationary anticyclone exists. Such anticyclones are prevalent in northern Europe in winter. In spring and summer they occasionally settle over the Caspian and Black Sea area, bringing cP air into Europe through the Balkans and the Ukraine.

MARITIME TROPICAL.–Due to the relatively low temperatures of the water areas southwest of Europe, the occurrence of maritime tropical (mT) air is not as extreme as in the United States.

The maritime tropical (mT) air of Europe originates south and east of the large semipermanent anticyclones of the Atlantic. The mT air then flows around the center of the anticyclone and enters Europe from the north or northwest.

CONTINENTAL TROPICAL.–European continental tropical (cT) air masses have their source regions in North Africa and in Asia Minor, where the air is unstable but very dry. These air masses appear in summer and winter, although they are rare in winter. As they move into Europe, much moisture is added and instability showers result. Continental air is confined to southern Europe, in the Mediterranean region.

Air Masses of Asia

As expected, over the vast land mass of Asia, the air masses exhibit even more extreme properties than those of North America. The coldest winter air masses (with the exception of Antarctica) and the warmest summer air masses are spawned in Asia. The area of most frequent cyclogenesis in the world is the Asiatic-Pacific area, where the extremely cold Arctic air in winter comes into contact with one of the warmest ocean areas. In summer, one of the foggiest regions in the world is where the warm air from the extensive land mass or from the large warm ocean areas to the southwest and south comes in contact with the cold waters of the Oyashio.

WINTER AIR MASSES.–The great winter air masses of Asia show the effects of more vertical mixing than do their counterparts in North America. This is due to the more general distribution of mountains in Asia and to the strong winds of the winter monsoon. The towering Himalayas and their eastern extensions prevent the continental air masses from mixing with the tropical air from the south; therefore, the Asian air masses are less variable than those in North America.

Similar to North America, the temperatures are lowest near the ground in the interior of the continent of Asia, but near the coast at comparable latitudes the temperatures are lower aloft.

The winter air masses usually have moved across vast areas of water before they encounter tropical air. On reaching the warm seas they become moisture laden and unstable, bringing to the lee side of mountains on the principal islands much shower activity. These air masses assume tropical characteristics by the time they reach the tropical seas.

SUMMER AIR MASSES.–As in winter, the air masses are similar to those in North America. In summer, the monsoon in Asia brings tropical air into the interior from the southeast, permitting considerable interaction of air masses. The tropical air that invades India and China in summer is extremely warm and moist. Some air invades India and the Far East from across the Equator from the Southern Hemisphere. The monsoon air may often be classified as equatorial.

Air Masses of the Southern Hemisphere

The air masses of the Southern Hemisphere are predominantly maritime. This is due to the overwhelming preponderance of ocean areas. Great meridional transports of air masses as they are known in the Northern Hemisphere are absent because the westerlies are much more developed in the Southern Hemisphere than in the Northern Hemisphere. Except for Antarctica, there are no large land masses in the high latitudes in the Southern Hemisphere; this prevents sizable invasions of Antarctic air masses. The large land masses near the Equator, on the other hand, permit the extensive development of warm air masses.

The maritime tropical air masses of the Southern Hemisphere are quite similar to their counterparts of the Northern Hemisphere. In the large area of Brazil, there are two air masses for consideration. One is the regular air mass from the Atlantic which is composed of unmodified mT air. The other originates in the Atlantic; but by the time it spreads over the huge Amazon River basin, it undergoes two important changes: the addition of heat and moisture. As a result of strong heating in summer, a warm dry air mass, continental tropical (cT) is located from 30° south to 40° south.

The maritime polar air that invades South America is quite similar to its counterpart in the United States. Maritime polar air occupies by far the most territory in the Southern Hemisphere, encircling it entirely and providing the momentum for the great west wind drift.

Australia is a source region for continental tropical air. It originates over the vast desert area in the interior. Except along the eastern coast, maritime tropical air does not invade Australia to a marked degree. This air is brought down from the north, particularly in the summer, by the counterclockwise circulation around the South Pacific high.

Antarctica is a great source region for intensely cold air masses. They have continental characteristics, but before the air reaches other land areas, it becomes modified and is properly called maritime polar. The temperatures are colder than in the Arctic regions. Results of Operation Deepfreeze have revealed the coldest surface temperatures for the world to be in the Antarctic.

During the polar night the absence of insolation causes a prolonged cooling of the snow surface which makes Antarctica a permanent source of very cold air. It is extremely dry and stable aloft. This polar air mass is referred to as Continental Antarctic Air, (cA).

In summer the continent is not as cold as in winter due to constant solar radiation but continues to function as a source for cold cA air.

In both winter and summer the air mass is thermally modified as it flows northward through downslope motion and surface heating and as a result becomes less stable. It assumes the characteristic of maritime Antarctic air. The leading edge of this air mass then becomes the northern boundary of the Antarctic front.

To the north of the Antarctic front is found a vast mass of maritime polar air which extends around the hemisphere between 40°S and 68°S in summer and between 34°S and 65°S in winter. At the northern limit of this air mass is found the Southern Hemisphere polar front. During summer this mP air is by far the most important cold air mass of the hemisphere due to the lack of massive outbreaks of cold continental air from Antarctica.

Different weather conditions occur with each type air mass.

1. The cA produces mostly clear skies.
2. The mA air masses are characterized generally by an extensive overcast of stratus and stratocumulus clouds with copious snow showers within the broad zone of the Antarctic front.
3. An area of transition which extends mainly from the coastline to the northern edge of the consolidated pack ice is characterized by broken to overcast stratocumulus clouds with somewhat higher bases and little precipitation.

FRONTAL CHARACTERISTICS

A FRONT is defined as the point of intersection of a frontal surface with a horizontal surface. A FRONTAL SURFACE is a surface of separation of two adjacent air masses of different densities, temperatures, and moisture content. It is taken to be the surface adjacent to the warm air. In reality, the discontinuity between adjacent air masses is not abrupt enough to warrant such a narrow definition. For this reason the concept of a FRONTAL ZONE was introduced. A FRONTAL ZONE is the transition zone between two adjacent air masses of different densities, bounded by a frontal surface. Since the temperature distribution is the most important regulator of atmospheric density, a front almost invariably separates air masses of different temperatures.

CONDITIONS NECESSARY FOR FRONTOGENESIS

Frontogenesis (the formation of a new front or the regeneration of an old one) takes place only when two conditions are met: (1) Two air masses of different densities exist adjacent to one another, and (2) a prevailing wind field exists, bringing them together.

There are three basic situations which are conducive to frontogenesis and which satisfy the two basic requirements. They are as follows:

1. The windflow is cross-isothermal as illustrated in figure 5-9, decreasing in speed downstream and flowing from cold air to warmer air. The flow need merely be cross-isothermal, not perpendicular, but the more perpendicular the cross-isothermal flow, the greater the intensity of frontogenesis.

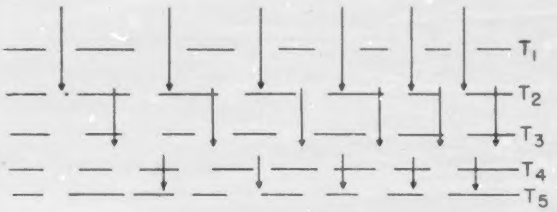

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Figure 5-9.—Frontogenetic cross-isothermal wind flow.

2. The winds of opposite air masses move toward the same point or line in a cross-isothermal flow. A classic example of this situation is the polar front where cold polar air moves southward toward warmer temperatures and warm tropical air moves northward toward colder temperatures. The boundary between them is the polar front.

3. The windstreams have formed a deformation field similar to the ones illustrated in figures 5-10 and 5-11.

A deformation field consists basically of a col area of flat pressure with two opposing highs and two opposing lows. It has two axes which have their origin at the neutral point in the col. The "y" axis, or axis of contraction, lies between the high and low which bring the air particles toward the neutral point. The "x" axis lies between the high and low which take air particles away from the neutral point and is known as the axis of dilation.

The distribution and concentration of isotherms in the deformation field determines whether frontogenesis will result. If the isotherms form a large angle with the axis of contraction, frontogenesis results; if a small angle, frontolysis results. It has been shown that in a perpendicular deformation field, isotherms must form an angle of 45° or less with the axis of dilation for frontogenesis to occur (fig. 5-10(A) and (B)). In a nonperpendicular deformation field, the critical angle changes correspondingly as illustrated in figure 5-11(A) and (B). In most cases, frontogenesis will occur along the axis of dilation. At any rate, frontogenesis will occur where there is a concentration of isotherms with the circulation to sustain that concentration.

FRONTOLYSIS

Frontolysis, or the dissipation of a front, occurs when either the temperature difference between the two air masses disappears or the wind carries the air particles of the air mass away from each other. Frontolytical processes are more common in the atmosphere than are frontogenetical processes. This comes about because there is no known property of the air which is conservative with respect to all the physical or dynamical processes of the atmosphere. The theory of frontogenesis is based on the assumption that there is such a property and that it is temperature. The nonconservative (principally nonadiabatic) influences on temperature must therefore be added to all the frontolytical processes.

Frontolytical processes are most effective in the lower layers of the atmosphere since surface heating and turbulent mixing are the most intense of the nonconservative influences on temperature.

It should be pointed out that for frontolysis to occur, only one of the two conditions stated above need be met. The simultaneous happening of both conditions results in more rapid frontolysis than if only one factor were operative.

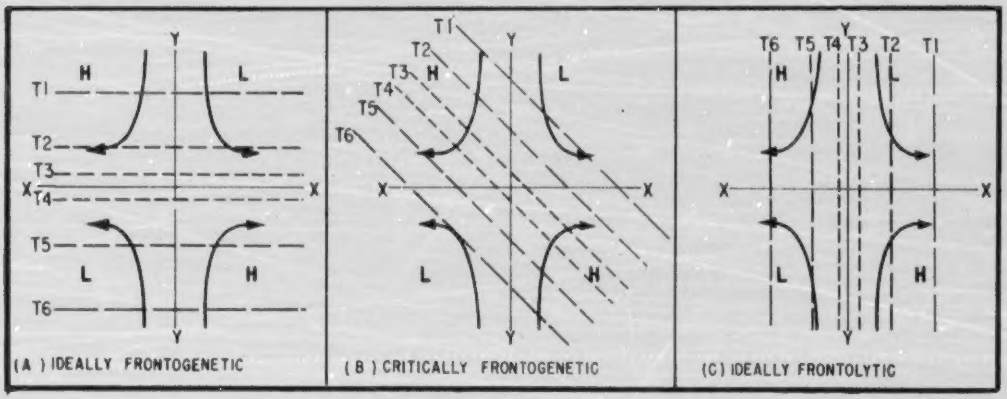
(A) IDEALLY FRONTOGENETIC
(B) CRITICALLY FRONTOGENETIC
(C) IDEALLY FRONTOLYTIC
T1
T2
T3
T4
T5
T6
H
L
X
Y

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Figure 5-10.–Perpendicular deformation field.

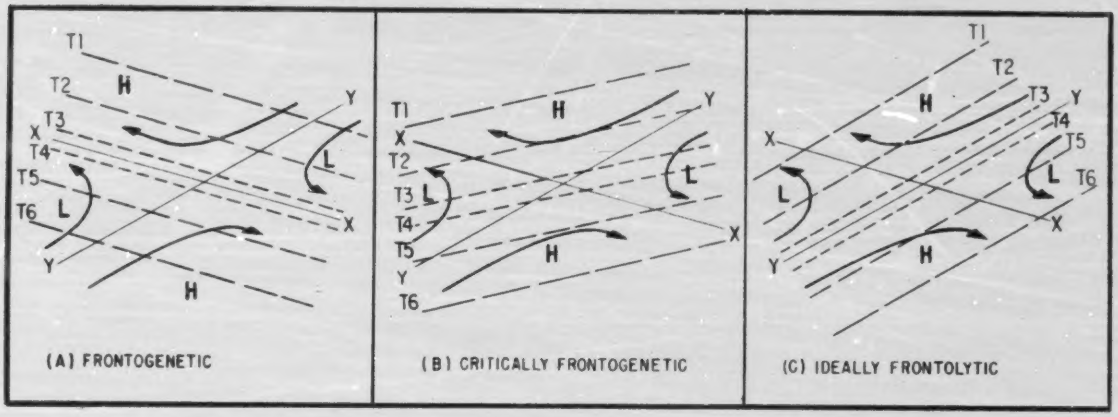
(A) FRONTOGENETIC
(B) CRITICALLY FRONTOGENETIC
(C) IDEALLY FRONTOLYTIC
T1
T2
T3
T4
T5
T6
H
L
X
Y

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Figure 5-11.–Nonperpendicular deformation field.

WORLD FRONTOGENETICAL ZONES

Certain regions of the world exhibit a high frequency for frontogenesis. These regions are coincident with the greatest temperature contrasts. Two of the most important frontal zones are those over the North Pacific and the North Atlantic Oceans as discussed in chapter 2 of this training manual. In winter, the Arctic front, a boundary between polar and Arctic air, forms in high latitudes over Northwest North America, the North Pacific, and near the Arctic Circle north of Europe. In summer, this front mainly disappears, except north of Europe.

The polar front, on the other hand, is present the year round, although not as intense in the summer as in the winter, due to a lessening temperature contrast between the opposing air masses. This front forms wherever the windflow and temperature contrast is favorable. Usually it is the boundary between tropical and polar air, but it may form between maritime polar and continental polar air. It also may exist between modified polar air and a fresh outbreak of polar air. This type is common over North America in the continental regions in winter in the vicinity of 50°N lat.

The polar front in winter is found most frequently off the eastern coasts of continents in the neighborhood of 30° to 60° latitude. It is also found over land; but since the temperature contrasts are greater between the continent and the oceans, especially in winter, the coastal areas are more favorable for formation and intensification.

The intertropical convergence zone, though not truly a front but a field of convergence between the opposing trades, forms a third semipermanent frontal type. This region shows a seasonal variation just as do the trades. The intertropical (or convergence zone) is thoroughly discussed in chapter 12 of this training manual.

SLOPE OF A FRONT

When we speak of the slope of a front, we are speaking basically of the steepness of the frontal surface, using a horizontal dimension and a vertical dimension. The vertical dimension used is normally one mile. A slope of 1:50 would be considered a steep slope and a slope of 1:300 a gradual slope. Factors favoring a steep slope are a large wind velocity difference between air masses, small temperature difference, and high latitude.

The frontal slope therefore depends on the latitude of the front, the wind speed, and the temperature difference between the air masses.

DISTRIBUTION OF ELEMENTS IN A FRONTAL ZONE

From our previous discussion and definitions of fronts, it was implied that a certain geometrical and meteorological consistency must exist between fronts at adjoining levels. It can also be inferred that the data at no one particular level is sufficient to locate a front with certainty in every case. In this section of the chapter the horizontal and vertical distribution of weather elements in a frontal zone will be discussed.

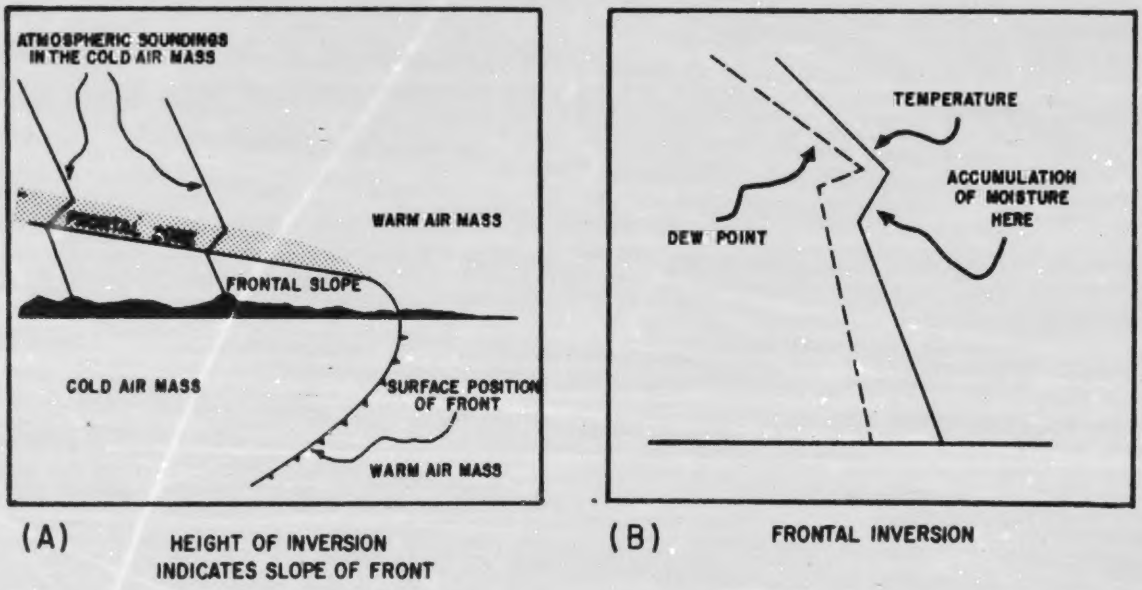
Temperatures

Since a front separates two different air masses, a frontal discontinuity should appear on a thermodynamic diagram as a straight line joining two different lapse rate curves. The top of the inversion or change in slope of the lapse rate would indicate the lower limit of the warm air, and the base of the inversion would indicate the depth of the cold air. The thickness of the frontal zone would be indicated by the thickness of the inversion layer. However, each air mass has characteristics which were acquired in its source region, but have been modified through changes caused by vertical motions, surface temperature influences, addition of moisture from evaporation of precipitation falling through the cold air mass, and other factors. Frontal zones therefore are often difficult to identify on upper air soundings.

The frontal discontinuity is not always an inversion of temperature. The degree to which the frontal zone appears pronounced is proportional to the temperature difference between two air masses.

Therefore, the primary indication of a frontal zone on a thermodynamic diagram is a decrease in the lapse rate somewhere in the sounding below 400 mb. The decrease in lapse rate may be a slightly less steep lapse rate for a stratum in a weak frontal zone to a very sharp inversion in strong fronts. In additon to a decrease in the lapse rate, there is usually an increase in moisture (a concurrent dewpoint inversion) at the frontal zone. This is especially true when the front is strong and abundant cloudiness and precipitation accompany it. Figure 5-12(A) shows the height of the inversion in two different parts of a frontal zone, and figure 5-12(B) shows a strong frontal inversion with a consequent dewpoint inversion.

The extent of the discontinuity, the strength of the inversion or isothermal layer, and the thickness of the zone of discontinuity depend on the discontinuity between the air masses; that is, the frontal intensity and the width of the frontal zone. Fronts are drawn on weather maps as sharp lines of discontinuity. This is done because of the scale of the charts. If the air masses were separated by a distinct line, the lapse rate


ATMOSPHERIC SOUNDINGS IN THE COLD AIR MASS
FRONTAL ZONE
WARM AIR MASS
FRONTAL SLOPE
COLD AIR MASS
SURFACE POSITION OF FRONT
WARM AIR MASS
TEMPERATURE
ACCUMULATION OF MOISTURE HERE
DEW POINT

(A) HEIGHT OF INVERSION INDICATES SLOPE OF FRONT

(B) FRONTAL INVERSION

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Figure 5-12.—(A) Height of inversion and thickness of inversion indicate frontal slope and intensity; (B) frontal inversion.

through the front would appear as shown in figure 5-13(A).

This lapse rate is considered the ideal lapse rate, but it is an obviously impossible one because two pressures and temperatures of different value cannot coexist at the same level side by side. Actual lapse rates through frontal zones therefore will lie between those as illustrated in figure 5-13(B), (C), and (D).

The edge of the frontal zone adjacent to the warm air is referred to as the frontal surface, and it is actually the intersection of this frontal surface with the weather map which is indicated at the line of discontinuity.

A cold front usually shows a marked temperature inversion. Also, higher relative humidity (dewpoint and mixing ratio) is indicated in both the air masses near the front.

In soundings through cold front transition zones, the dewpoint ordinarily is lower in the upper air mass if that air mass has the larger

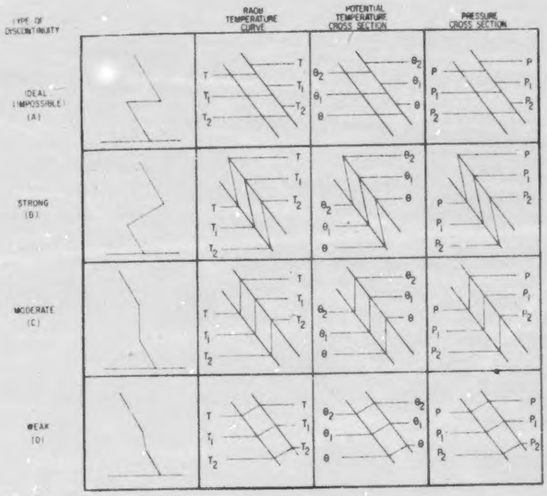
TYPE OF DISCONTINUITY
RAOB TEMPERATURE CURVE
POTENTIAL TEMPERATURE CROSS SECTION
PRESSURE CROSS SECTION
IDEAL (IMPOSSIBLE) (A)
STRONG (B)
MODERATE (C)
WEAK (D)
T
T1
T2
θ
θ1
θ2
P
P1
P2

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Figure 5-13.—Discontinuities through a frontal zone.

horizontal wind component toward the surface position of the front and therefore has a new downward motion.

The air above the cold front surface has an upward component when the horizontal wind component normal to the frontal surface decreases with height through the transition zone. In these cases, the dewpoint usually increases upward through a relatively well-marked transition zone.

Cold fronts generally show a stronger inversion than warm fronts, and the inversion will appear at successively higher levels as the front moves past a station. The reverse is true of warm fronts. Occluded fronts generally show a double inversion. However, as the occlusion process continues, an amalgamation of the air masses takes place, and the inversions are wiped out, or they fuse.

It is very important in raob analysis not to confuse the subsidence inversion of polar and Arctic air masses with frontal inversions. Extremely cold continental Arctic air, for instance, has a strong inversion which extends to the 700-mb level.

Sometimes it is difficult to find a significant stable zone on a particular sounding, though it is known that a front intersects the column of air over a given station. This may be because of adiabatic warming of the descending cold air just under the frontal surface or excessive local vertical mixing in the vicinity of the frontal zone. Under conditions of subsidence of the cold air beneath the frontal surface the subsidence inversion within the cold air may be more marked than the frontal zone itself. Vertical wind shear criteria then become even more important in the analysis.

Sometimes fronts on a raob sounding which might show a strong inversion often are accompanied by little weather activity. This is due to subsidence in the warm air, which will strengthen the inversion. The activity at the front increases only when there is a net upward vertical motion of the warm air mass.

Potential Temperature

Potential temperature is a conservative air mass property.

If the potential temperature of a frontal surface has been determined from the soundings and no representative temperatures that high can be found at the ground in the area where the frontal surface should be expected to intersect the ground, the front is designated on the surface chart as an "upper front" if the stratum of air over which this upper front lies is 1,500 feet or more in thickness. True fronts, therefore, would not be crossed by potential temperature lines. In nature, however, true discontinuities do not occur; mixing takes place in the frontal zone. There is usually a crowding of potential temperature lines at fronts.

The potential temperature of the top of the frontal zone is practically independent of elevation, except for a slight increase at higher levels. For example, in the cooler season the potential temperature (θ) of the polar front may increase from 298°K near the surface to 302°K at the 500-mb level. The θ of the frontal surface may also vary in the horizontal, depending upon the distribution of θ in the warm air mass. Fronts tend to have their highest potential temperatures in the Southwest United States, when they are next to continental tropical air which has a θ of about 308° to 312°K in the warmer seasons. Fronts in contact with maritime tropical air (polar fronts) have lower values of potential temperature, about 300°K. During the winter months the θ of the polar front is about 298°K. In the warmer season the polar front is found at a θ of 302°K. The Arctic front is generally found at a potential temperature of about 286°K.

Wind

Since winds near the earth's surface blow mainly along the isobars with a slight drift toward lower pressure, it follows that the wind direction in the vicinity of a front must conform with the refraction of the isobars. The arrows in figure 5-14 indicate the winds that correspond to the pressure distribution.

From this it can be seen that a front is a WIND SHIFT LINE and that wind shifts in a cyclonic direction. Since the front moves in the direction of the wind component normal to the front (heavy arrow on diagram), we can evolve the following rule: IF YOU STAND WITH

YOUR BACK AGAINST THE WIND IN ADVANCE OF THE FRONT, THE WIND WILL SHIFT CLOCKWISE (VEER) AS THE FRONT PASSES.

The speed of the wind depends upon the pressure gradient. Thus, in figure 5-14(A) the speed will be about the same in both air masses; in (B) and (C) a relatively strong wind is followed by a weaker wind, and in (D) a weak wind is followed by a strong wind.

An essential characteristic of a frontal zone is a wind discontinuity through the zone. The wind normally increases or decreases in speed with height through a frontal discontinuity. Backing usually occurs through a cold front and veering through a warm front. The sharpness of the wind discontinuity is proportional to the temperature contrast across the front and the pressure field in the vicinity of the front (the degree of convergence between the two air streams). With the pressure field constant, the sharpness of the frontal zone is proportional to the temperature discontinuity (no temperature discontinuity–no front; thus, no wind discontinuity). The classical picture of the variation in wind along the vertical through a frontal zone is shown in figure 5-15.

An example of a frontal zone and the winds through the frontal zone is shown in figure 5-16.

On this sounding the upper winds above the surface layer which show the greatest variation are those in the 800- to 650-mb layer. This indication coincides closely with the frontal indications of the temperature (T) and dewpoint (T_d) curves. Since the wind veers with height through the layer, the front would be the warm type.

The thermal wind can also be valuable indicator for the location of frontal zones. The magnitude of the thermal wind for a layer indicates the strength of the horizontal gradient of mean temperature of the layer. Since a frontal zone defines a layer of maximum horizontal thermal gradient, it also represents a zone of maximum thermal wind. Furthermore, since the thermal wind blows parallel to the mean isotherms for the layer (with cold air to the left), the direction of the thermal wind vector is roughly representative of the directional orientation of the front. The winds aloft for the sounding shown in figure 5-16 are seen plotted as a hodograph in figure 5-17.

The thermal wind is indicated by the shear vector connecting the wind vectors for the base and top of each layer. The maximum thermal wind is found between 700 and 650 mb, supporting the previous frontal indications. The direction of the thermal wind vector from 800 to 650 mb is roughly north-south; the frontal zone could thus be assumed to be oriented along a generally north-south line.

In conclusion, the vertical wind shift through a frontal zone depends on the direction of the slope. In cold fronts the wind backs with height, and cold advection is therefore indicated. In warm fronts the wind veers with height, indicating warm air advection. At the surface, the wind ALWAYS veers across the front and the isobars have a cyclonic kink which points toward higher pressure. Sometimes the associated pressure trough is not coincident with the front; in such

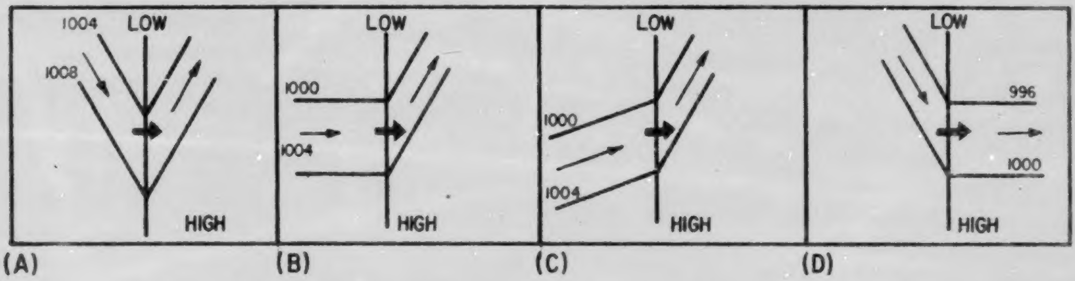

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Figure 5-14.—Types of isobars associated with fronts.

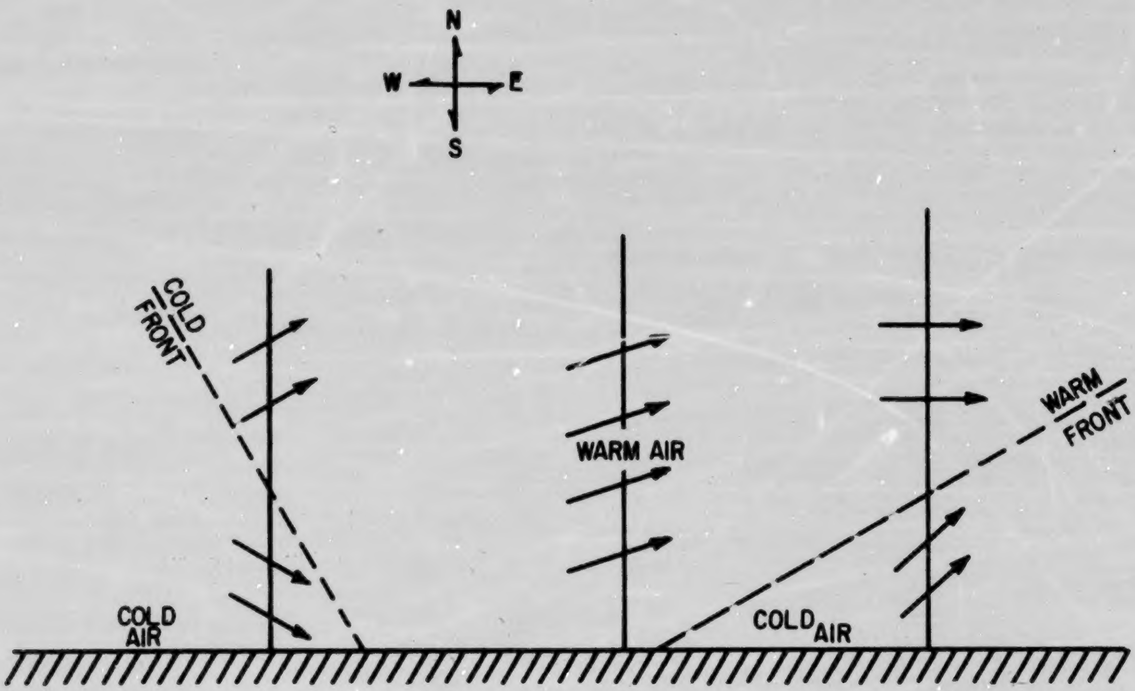

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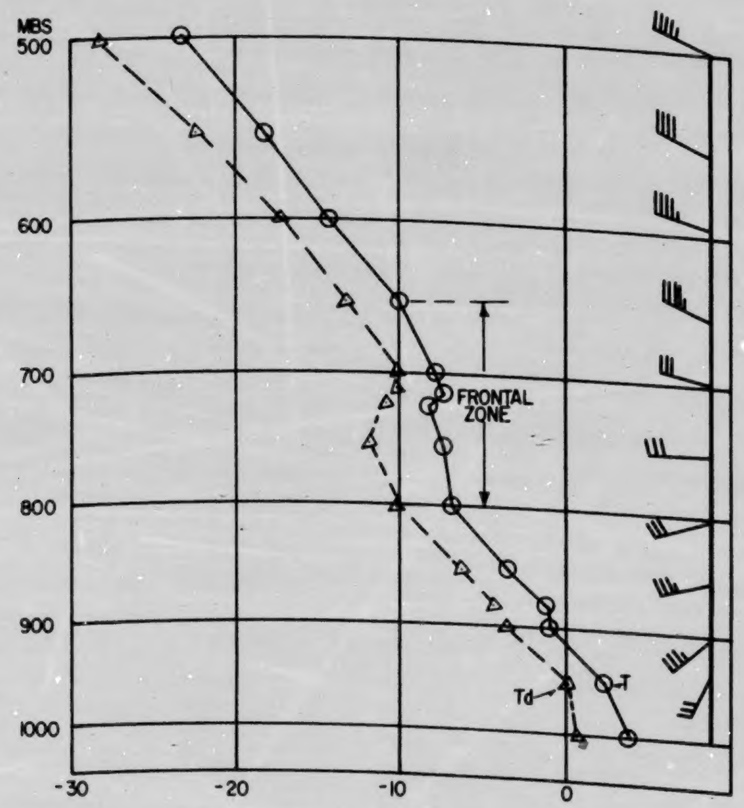
Figure 5-15.—Vertical distribution of wind direction in the vicinity of frontal surfaces.

cases there may not be an appreciable wind shift across the front—only a speed discontinuity. When the lowest pressure in the associated trough is found ahead of the front, the winds are of the same direction, but the speed is greater BEHIND the front. When the lowest pressure in the associated trough is found behind the front, the wind direction is the same on either side of the front, but the wind speed is greater AHEAD of the front. The reason for all of the foregoing is that a front must have a cyclonic shear across it to be a front. When the cyclonic shear disappears, so does the front. The reasoning behind this cyclonic shear requirement can be found in Margules equation of frontal slopes. The direction of the slope is determined by the wind speed difference between the cold air and the warm air. The sign of the difference determines the direction of the slope with the qualification that the slope must be toward the cold air. The potentially warmer air always lies over the potentially colder air. This is also demonstrated by actual upper air soundings where it will be found that a specific potential temperature is always found at a higher altitude in the cold air than in the warm air. Stated in a different manner, the potential temperature in the warm air is higher, level for level.

Clouds and Weather

Cloud decks are usually in the warm air mass due to the upward vertical movement of the warm air. Clouds forming in the cold air mass are due to evaporation of moisture from precipitation from the overlying warm air mass and/or by vertical lifting. Convergence at the front results in a lifting of both types of air. The stability of the air masses determines the cloud and weather structure at the fronts as well as the weather in advance of the fronts.



igure 5-16.–Distribution of wind and temperature through a warm frontal zone.

CLASSIFICATION OF FRONTS

The frontal characteristics described so far are general characteristics of fronts and may be applied as to whether cold air replaces warmer air or warm air replaces colder air. Some specific cases were mentioned to illustrate frontal characteristics. However, depending upon the movement of the front and the stability conditions of the air masses, a number of additional characteristics must be considered. Fronts are classified as follows (motion relative to the warm and cold air masses is the criterion):

1. COLD FRONT. A cold front is one that moves in a direction in which cold air displaces warm air at the surface.
2. WARM FRONT. A warm front is one along which warmer air replaces colder air.
3. QUASI-STATIONARY FRONT. This type front is one along which one air mass does not appreciably replace the other.
4. OCCLUDED FRONT. An occluded front is one where the cold front overtakes the warm front and the warm air is squeezed upward. The occluded front may be either a WARM FRONT TYPE, one in which the cool air behind the cold front overrides the colder air in advance of the warm front, resulting in a cold front aloft; or a COLD FRONT TYPE, one in which the cold air behind the cold front underrides the warm front and the warm front is aloft.

FRONTAL TYPES

Whether a front is to be classed as quasi-stationary, warm, or cold is determined on the

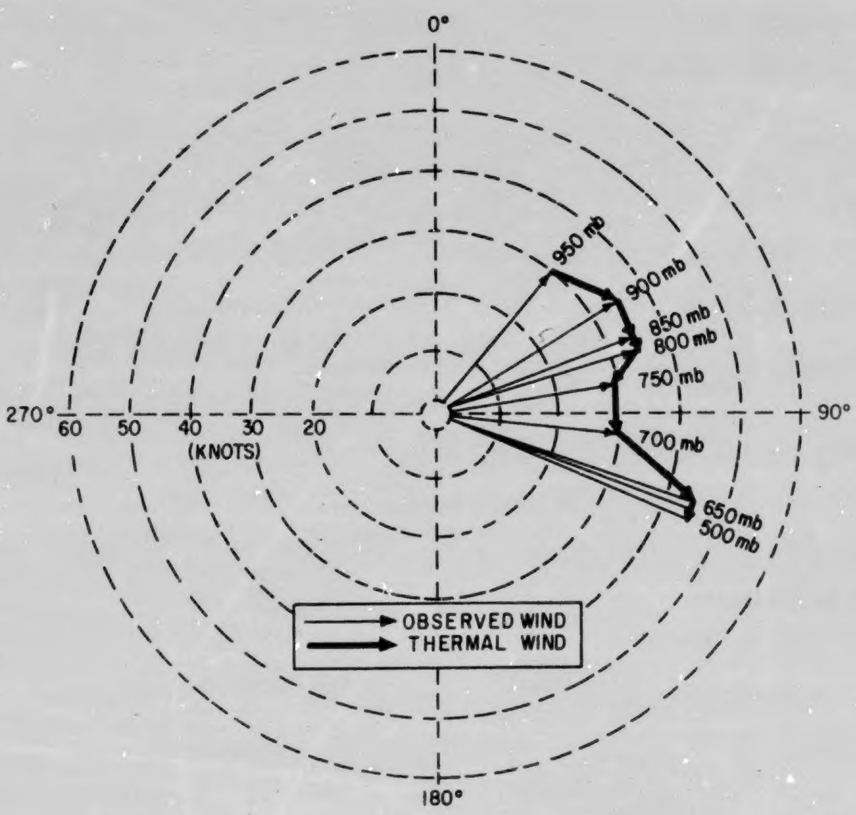


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Figure 5-17.—Hodograph of observed and thermal winds for sounding in figure 5-16.

basis of the instantaneous or known direction of movement. Past movements for 3 to 6 hours may be used as a guide. The instantaneous direction of motion is determined insofar as possible on the basis of actual winds in the cold air rather than on geostrophic winds. If representative wind observations in the cold air are not available, the direction of the front can be inferred from the pressure gradient and past history of evidence of passage of the front at individual stations.

If isobars intersect the front with low pressure to the north (westerly winds) and colder air is to the west, the front is usually a cold front. The following examples illustrate this point:

1. If isobars with easterly type flow are cyclonic or straight in the colder air mass, the designation will usually be a warm front, because cyclonic curvature means ordinarily that there is a component of motion of the cold air away from the front. In this case the direction of motion of the front is consistent with that indicated by the pressure gradient along the front. (See fig. 5-18.)

2. If isobars with easterly type flow are anticyclonic in the colder air mass, the designation will usually be cold or quasi-stationary, as illustrated in figure 5-19.

This case occurs only when the pressure gradient along the front is slight, or the isobars cross the front only at relatively long intervals of distance.

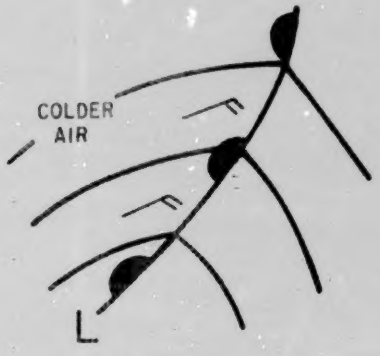

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Figure 5-18.—Warm front in an easterly flow.

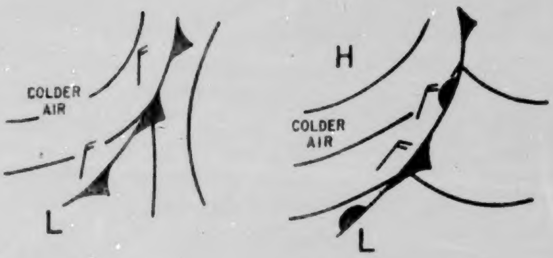

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Figure 5-19.—Cold and quasi-stationary fronts in an easterly flow.

Warm Fronts

Normally, the eastern part of the frontal system of any nonoccluded cyclone or open wave is properly designated as a warm front, but there may be some cases of open waves where caution is necessary because high-pressure conditions on the cold air side may prevent movement of the front; the designation such cases is usually quasi-stationary.

Cold Fronts

A surface front is indicated as "cold" if the low-level wind direction in the cold air along any appreciable segments of the front has a component, however small, toward the front or into the warm air. If the cold air wind component toward the front is very small, the quasi-stationary character of the front may be indicated.

A problem sometimes arises during the winter seasons when such a cold wind component toward the warm air exists; but because of daytime heating, the leading edge is continually destroyed and the front appears to remain stationary and may even retrograde during the day. The policy is to continue to indicate this front as cold, since the cold air component is toward the warm air. Actually, the cold air will make progress toward the warm at night, in keeping with its designation as a cold front. This condition is encountered especially with mP fronts from the Pacific moving into the interior of the Pacific Northwest States in summer.

The only criterion for classification of a front as cold is its direction of motion, requiring that the component of true horizontal surface wind in the cold air, perpendicular to the front and adjacent to it, be directed toward the warm air. The true wind may be different from the corresponding component of the geostrophic wind, as illustrated in figure 5-19. With southward moving cold fronts along the east slope of the Rockies, true winds in the cold air are sometimes nearly perpendicular to both the front and the sea level isobars.

Occluded Fronts

The Aerographer's Mate should designate as "occluded fronts" only those fronts which result from the classic occlusion process that takes place when an unstable wave develops along what was originally a single front.

For example, if a cold type occlusion overtakes a warm or stationary front along the north Pacific coast and is forced aloft, the coastal warm front overtaken is not then indicated as a warm type occlusion because it is not the result of a normal occlusion process from an unstable wave. There are several reasons for this convention, one being that the process illustrated by this example is usually not accompanied by cyclonic development, and another that the potential temperatures along the two frontal surfaces may be widely different; whereas, with the normal occlusion process, the temperatures along the two fronts are initially about the same.

A situation similar to the overtaking of a coastal warm front occurs when maritime polar air from the Pacific overtakes a quasi-stationary Arctic front along the eastern slope of the Continental Divide. In this case the best practice is to continue to designate the portion of the Arctic front overtaken by the new surge of mP air as quasi-stationary, cold, or warm, as appropriate, and not as a warm type occlusion. In this case the two fronts have markedly different potential temperatures, and it is possible under proper circumstances for a wave to then form along the Arctic front and progress to the occluded state.

WARM OCCLUSIONS.–A warm occlusion, or warm-front type occluded front, develops when the temperature of the leading edge of the cold front, as it overtakes the warm front, is higher than that of the warm front at the point of contact. The term "warm occlusion" is applied to that portion of the surface warm front which has been overtaken by the cold front, provided the process is that of an unstable occluding wave cyclone. The warm occlusion usually develops in the Northern Hemisphere when conditions to the north of the warm front, such as the existence of an anticyclone, maintain low temperatures north of the warm front while the trajectory of the cold air behind the cold front is such as to cause warming. The cold front, in this case, continues as an upper cold front above the warm front surface.

A special case arises when the center of a low having a warm type occlusion moves along the occlusion or when a new low develops along the occlusion or at the apex of the warm sector. When this happens, the original warm type occlusion necessarily reverses direction and begins to move toward the west or south as a cold front. In order to avoid confusion, since the front no longer moves as a warm front and is not by history a cold type occlusion, the front should be redesignated as a "simple cold front."

COLD OCCLUSIONS.–A cold occlusion, or cold-front type occluded front, develops when the temperature of the leading edge of the cold front, as it overtakes the warm front, is lower than that of the warm front as the point of contact. The term "cold occlusion" is applied to that portion of the surface cold front which has overtaken the surface warm front, provided the process is that of an unstable occluding wave cyclone. Cold occlusions are more frequent than warm occlusions. The lifting of the warm front as it is underrun by the cold front implies existence of an upper warm front to the rear of the cold occlusion; actually such a warm front aloft is rarely discernible or significant.

Most fronts approaching the Pacific coast of North America from the west are cold occlusions. In winter these fronts usually encounter a shallow layer of surface air near the coastline, from about Oregon northward, that is colder than the leading edge of cold air to the rear of the occlusion. Sometimes, especially off British Columbia, this coastal layer of cold air extends far enough outward from the coastline to warrant its delineation by a surface stationary front. The usual practice in these cases is to continue to designate the cold occlusion as though it were a surface front because of the shallowness of the layer over which it rides. Ordinarily the portion of the coastal stationary front which has been overtaken by the occlusion may be dropped from the analysis and should be completely dropped after the entire occlusion has moved inland. The coastal stationary front itself is prevented by topographic features from moving inland, at least beyond the higher mountains which are generally within 100 miles of the coastline. The passage of the cold type occlusion over the coastal layer of colder air presents a difficult problem of analysis in that no surface wind shift will ordinarily occur at the exact time of passage. However, a line of stations reporting surface pressure rises is the best criterion of its passage. This should be verified by reference to plotted raob soundings where available. When a Pacific cold occlusion moves farther inland, it may encounter colder air of appreciable depth over the Plateau or Western Plains areas, in which case it should be redesignated as an "upper cold front."

Quasi-Stationary Fronts

A front is classed as "quasi-stationary" if there is no movement of cold air normal to it or if the amount of movement is too small for its direction to be determined, as may be the case with minor undulations along the front.

The quasi-stationary designation tends to be used too frequently. Many of the fronts so designated move slowly but steadily in one direction or the other. For example, a front moving 10 miles per hour between stations 70 miles apart could be found on three successive 3-hourly synoptic maps still between the same stations. Such fronts are actually slowly moving cold fronts or slowly moving warm fronts, and should be thus designated. Fronts which are often designated "quasi-stationary" by successive analysts on many map sequences may be found to have traversed great distances in 12 to 24 hours, not infrequently 200 to 300 miles.

A special type of quasi-stationary front which is frequently very well marked by wind and dewpoint discontinuity as the surface is the so-called dewpoint front extending north-south usually through Texas and Oklahoma; for example, during the warmer months. This line of demarcation between hot cT to the west and cooler mT air to the east either slopes upward to the east or is nearly horizontal. The hot dry cT air of the Southwest United States is therefore usually continuous with warm dry air aloft over the Central Plains area and southern United States. Since this discontinuity rarely moves east of about the longitude of Dallas, Texas, and is a climatological feature of the region, Analysis Centers at times unfortunately do not designate it as a front.

FRONTAL INTENSITY

No completely acceptable set of criteria is in existence as to the determination of frontal intensity, as it depends upon a number of variables. Some of the criteria which may be helpful in delineating frontal intensity are discussed in the following paragraphs.

Turbulence

Except when turbulence or gustiness may result, weather phenomena are not taken into account when specifying frontal intensity, because a front is not defined in terms of weather. A front may be intense in terms of discontinuity of density across it, but may be accompanied by no weather phenomena other than strong winds and a drop in temperature. A front which would otherwise be classified as weak is considered moderate if turbulence and gustiness are prevalent along it, and an otherwise moderate front is classified as strong. The term gustiness for this purpose is taken to include convective phenomena such as thunderstorms and strong winds regardless of the amount of wind shear.

Temperature Gradient

Temperature gradient, rather than true difference of temperature across the frontal surface, is used in defining the frontal intensity in terms of temperature. The temperature difference between the two air masses does not occur across a surface of zero thickness as required for the true definition of a front. Rather, the difference is distributed across a transition zone which is usually some 25 to 50 miles in width. In addition, there is normally a temperature gradient away from the front in the cold air mass.

Occasionally, autographic records of frontal passages indicate that in addition to the transition zone there is sometimes a change of temperature at the instant of arrival of the cold air, which suggests a nearly perfect discontinuity of density across the front. This is usually in addition to a greater net decrease of temperature across the transition zone.

Temperature gradient, when determining frontal intensity, is defined as the difference between the representative warm air immediately adjacent to the front and the represetative surface temperature 100 miles from the front on the cold air side. By convention, the transition zone is taken to be part of the cold air mass.

A suggested set of criteria based on the horizontal temperature gradient just described has been devised. It defines a weak front as one where the temperature gradient is less than 10°F per 100 miles; a moderate front where the temperature gradient is 10° to 20°F per 100 miles; and a strong front where the gradient is over 20°F per 100 miles.

The 850-mb level temperatures may be used in lieu of the surface temperatures if representative surface temperatures are not available and the terrain elevation is not over 3,000 feet. Over much of the western section of the United

States, the 700-mb level temperatures can be used in lieu of the surface temperatures.

WIND SHEAR

The wind shear may be either the vector difference between components of surface geostrophic wind parallel to and immediately on either side of the front or the 1,000-500 mb thermal wind shear.

Surface Wind Shear

The wind contrast (shear) is normally taken as the vector difference between the components of surface geostrophic wind parallel to and immediately on either side of the front, as illustrated in figure 5-20.

"W" and "C" in the figure are vectors representing the computed geostrophic wind in the warm and cold air, respectively; and the "w" and "c" are the respective components parallel to the front. In (A), the wind contrast is w − (−c) = w + c, and in (B) it is w − c.

Thermal Wind Shear

The two most important properties ascribed to the frontal boundries by the National Meteorological Center (NMC) are a pressure gradient discontinuity and a three-dimensional temperature gradient discontinuity. The frontal intensity criteria are based directly upon the intensity of the thickness gradient discontinuity (to be discussed in a later section) and upon weather along the front from which intensity is taken as a measure of the convergence.

In actual practice, the average thermal wind over a 5-degree band on both sides of the front is determined by averaging the thermal winds or computing them from the thickness gradient. The thermal wind shear is converted into frontal intensity using the following relationships:

If the thermal wind shear is equal to or less than 25 knots, no front exists or frontolysis is in progress; if the thermal wind shear is greater than 25 knots, but equal to or less than 50 knots, it is a weak front or frontogenesis is in progress; if the thermal wind shear is greater than 50 knots but equal to or less than 75 knots, the front is of moderate intensity; and if the thermal wind shear is over 75 knots, the front is strong.

NMC also modifies the intensities of fronts on the basis of observed weather associated with the front. Clear skies and weak or no surface wind shift decrease the intensity one category. Moderate to heavy precipitation or pronounced surface wind shifts increase the intensity one category.

POLAR FRONT THEORY

The polar regions are dominated by cold air masses and the Tropics by warm air masses. The middle latitudes are regions where cold and warm air masses continually interact with each other, the cold air moving southward and the

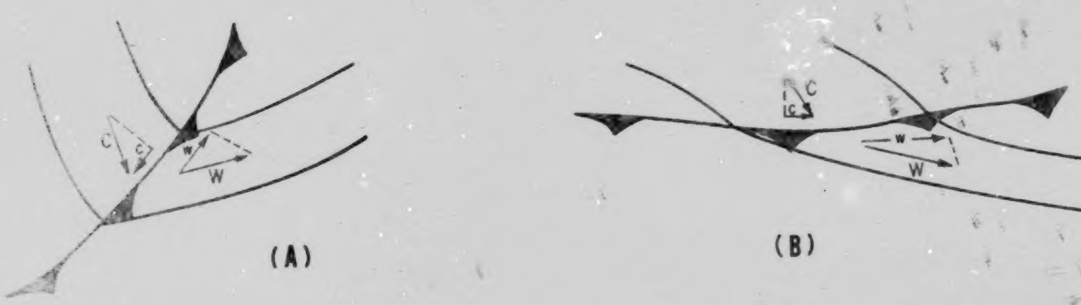

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Figure 5-20.—Determination of wind shear across a front for determination of frontal intensity. (A) Wind components in opposite direction; (B) wind components parallel to front in the same direction.

warm air moving northward in alternating tongues or waves. The zone which separates these air masses is the polar front. When conditions are favorable, extratropical cyclones develop on the polar front. They form as wavelike perturbations on the front and go through a life cycle. This cycle can be either of two patterns, that of stable waves or unstable waves.

STABLE WAVES

A stable wave is one that neither develops nor occludes, but appears to remain in about the same state. Stable waves are usually of small amplitude and have a fairly regular rate and direction of movement.

UNSTABLE WAVES

This type wave is by far the more common that is experienced with development along the polar front. The amplitude of this wave increases with time until the occlusion process sets in; and when the occlusion process is complete, the polar front is reestablished.

Figure 5-21 illustrates the development and occlusion of an unstable wave cyclone.

In its initial stage of development the polar front separates the polar easterlies from the midlatitude westerlies (A); the small disturbance due to the steady state of the wind is often not obvious on the weather map. Uneven local heating, irregular terrain, or wind shear between the opposing air currents may start a wavelike perturbation on the front (B); if this tendency persists and the wave increases in amplitude, a counterclockwise (cyclonic) circulation is set up. One section of the front begins to move as a warm front while the adjacent sections begin to move as a cold front (C). This deformation is called a frontal wave.

The pressure at the peak of the frontal wave falls, and a low-pressure center is formed. The cyclonic circulation becomes stronger, and the wind components are now strong enough to move the fronts, the westerlies turn to southwest winds and push the eastern part of the front northward as a warm front, and the easterlies on the western side turn to northerly winds and push the western part southward as a cold front. The cold front is moving faster than the warm front (D). When the cold front overtakes the warm front and closes the warm sector, an occlusion is formed (E). This is the time of maximum intensity of the wave cyclone.

As the occlusion continues to extend outward, the cyclonic circulation diminishes in intensity (the low-pressure area weakens), and the frontal movement slows down (F). Sometimes a new frontal wave may now begin to form on the westward trailing portion of the cold front. In the final stage, the two fronts become a single stationary front again. The low center with its remnant of the occlusion has disappeared (G).

CYCLONES

DEFINITIONS AND TERMINOLOGY

The term cyclone is used to denote any area of closed counterclockwise circulation in the Northern Hemisphere. Because of the basic wind-pressure relationships, the center of such an area will also be a relative minimum in the pressure field. The term low is used interchangeably with the term cyclone. A distinction is also made between those cyclones forming outside the Tropics (extratropical) and those forming in the Tropics (tropical). The latter are discussed in chapter 12 of this training manual. In this section the word cyclone applies only to those forming outside the Tropics. If the central pressure of a cyclone is decreasing with time, it is said to be deepening; filling denotes the converse situation. An unstable wave cyclone is usually developing, whereas a stable wave is not. When the more rapidly moving cold front of a wave cyclone overtakes the warm front, the cyclone is said to be an occluded cyclone. The typical stages of development of an occluded wave cyclone were illustrated in the previous section.

CYCLOGENESIS

There is a systematic relationship between cyclones and fronts, in that the cyclones are usually associated with waves along fronts–primary cold fronts. Cyclones come into being or intensify because pressure falls more rapidly

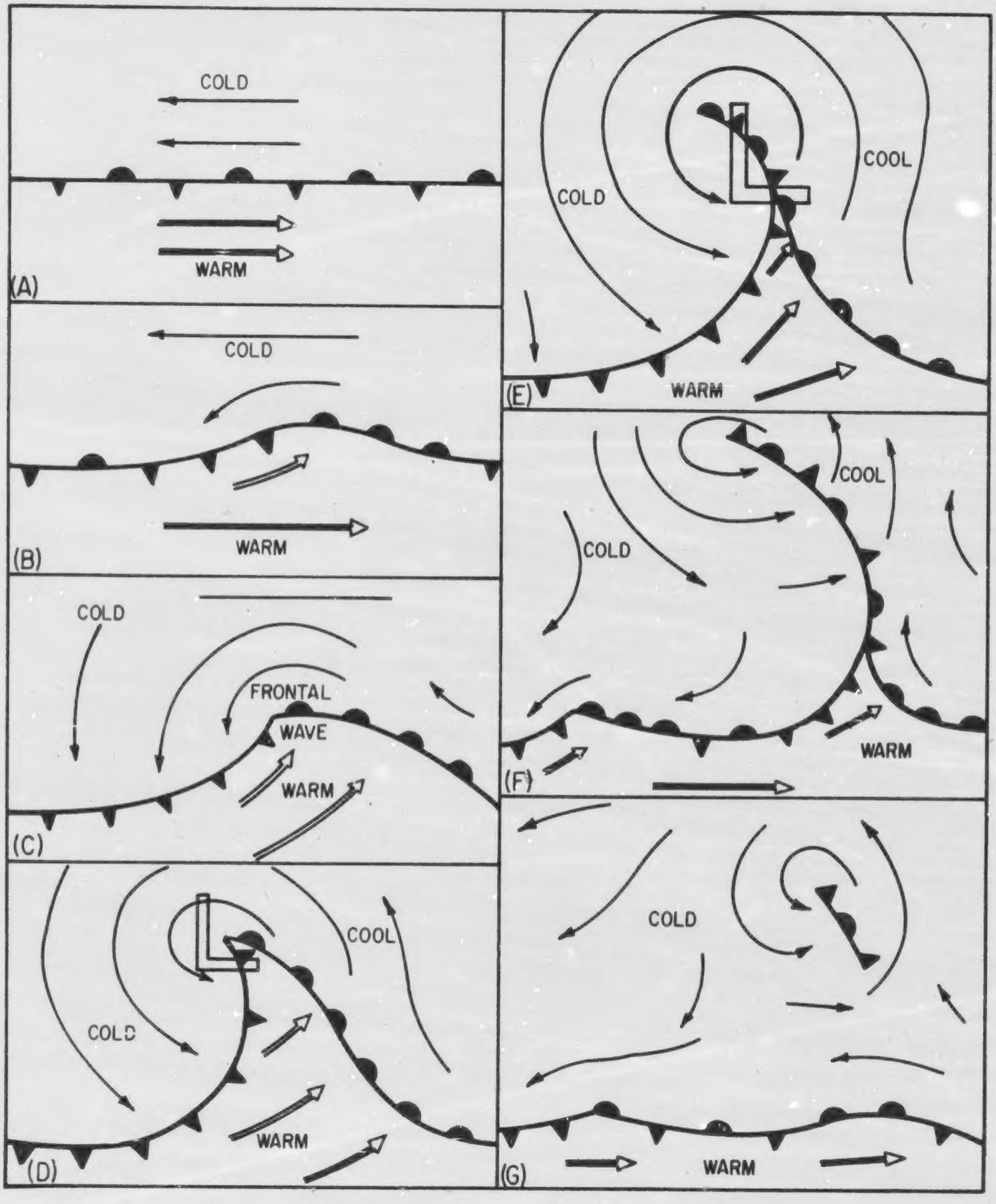
COLD
WARM
(A)
COLD
WARM
(B)
COLD
FRONTAL
WAVE
WARM
(C)
COLD
COOL
WARM
(D)
COLD
COOL
WARM
(E)
COLD
COOL
WARM
(F)
COLD
WARM
(G)

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Figure 5-21.—The life cycle of an unstable frontal wave.

at one point than it does in the surrounding area. The resultant increase in the intensity of the counterclockwise circulation (Northern Hemisphere) is called cyclogenesis. It can occur anywhere, but in middle and high latitudes it is most likely to occur on a frontal trough. In this instance, the cyclogenesis begins at the lowest level and works gradually upward as the cyclone deepens. The reverse also occurs; closed circulations aloft sometimes work downward until they appear on the surface chart. These cyclones rarely contain fronts, and are quasi-stationary or drift slowly westward and/or equatorward. The surface process however is usually a combination of two factors: perturbations along the polar front in conjunction with atmospheric waves aloft.

Wave cyclones, as opposed to those which work downward from upper air charts, normally progress along the polar front with an eastward component at an average rate of 25 to 30 knots, although 50 knots is not impossible, especially in the case of stable waves. Both the speed and direction of movement of the surface cyclone provide good clues as to the nature and vertical extent of cyclogenesis. It should be noted here that cyclogenesis and deepening are not synonymous terms. Deepening very often accompanies cyclogenesis, but not necessarily. For example, a certain amount of deepening or filling results solely from the diurnal variation of pressure.

In considering the likelihood of cyclogenesis, the forecaster should keep in mind the basic definition of a relative minimum in the pressure field as the intersection of two or more troughs. Whenever a faster moving upper trough overtakes a surface front and its associated trough, cyclogenesis is likely. This is an especially important occurrence when a progressive short-wave trough from the Rockies moves over a quasi-stationary front along the Gulf Coast in winter.

Wave formation is more likely on slowly moving or stationary fronts than on rapidly moving fronts, and certain oragraphical areas are preferred localities for cyclogenesis. The Rockies, the Ozarks, and the Appalachians are examples in North America.

The most common indications of wave cyclogenesis are:

1. Marked increase in katallobaric gradients, (falling pressure) especially in the cold air.
2. Marked location changes in wind direction or speed.
3. Increasing cloudiness north of the frontal area.
4. Start of precipitation on the cold air side of the front.

The above criteria do not always give adequate warning; in many cases the cyclone has already formed by the time these changes appear on the surface data. The upper air indications must be taken into account with simultaneous occurrences on the surface.

The shape and curvature of the isobars also give valuable indications of frontogenesis and therefore possible cyclogenesis.

On a cold front anticyclonically curved isobars behind the front indicate that the front is slow moving and therefore exposed to frontogenesis. Cyclonically curved isobars in the cold air behind the cold front indicate that the front is fast moving and exposed to frontolysis.

On the warm fronts the converse is true. Anticyclonically curved isobars in advance of the warm front indicate the front is fast moving and exposed to frontolysis and with cyclonically curved isobars the warm front is retarded and exposed to frontogenesis. Figure 5-22 illustrates the frontogenetical conditions around an ideal wave. FG denotes frontogenesis and FL frontolysis.

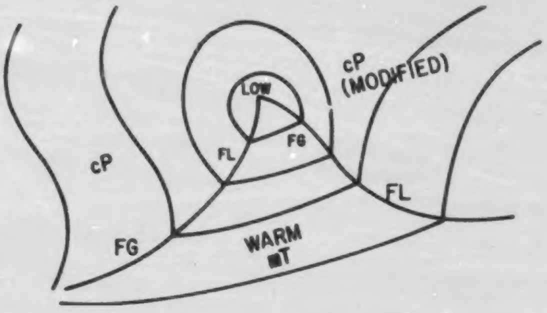

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Figure 5-22.—Ideal wave with indications of frontogenesis and frontolysis from the curvature of the isobars.

Some of the major causes for perturbations along fronts, that is, the formation of waves, are:

1. Disturbance in the nearby westerlies aloft. A cyclonic flow or higher wind speeds in the cold air in the westerlies aloft produce a cyclonic shear and therefore tend to set up a cyclonic circulation which builds downward to the surface. The surface low thus formed will continue to develop and exist as long as it is in association with this disturbance in the westerlies aloft and a depletion of mass or divergence is taking place in the system.

2. Superposition of a jet maximum over a front. Due to the steep gradient of temperature in the cold air aloft, wind speed gradients are also steeper in the cold air aloft and the circulation takes on a strong cyclonic shear. This shear in turn leads to a cyclonic circulation aloft which will build downward to the surface. The introduction of the cyclonic circulation in the front causes formation of waves along the front and marks the beginning of the life cycle of the cyclone. The mechanism for the reduction of pressure in the center of the cyclone is identical with that for disturbances in the westerlies. Some meteorologists make no distinction between these two causes of cyclogenesis because of their identical function in cyclogenesis.

3. Mountain waves. These mountain waves often generated to the lee of the Rocky Mountains and the Appalachian Mountains can set up cyclonic circulations aloft independent of other influences. Mountain waves may be only influential in weakening a high, generating an independent cyclone, or, as is more often the case, generating a cyclonic circulation on a front in their proximity. The Texas and Colorado lows are prime examples of the role played by mountain waves in cyclogenesis along fronts. When an independent cyclone is generated, it soon draws nearby fronts into its circulation.

4. Coastlines. Rapid changes in friction acting on the wind can deflect the wind in direction and speed and initiate a cyclonic circulation. Sudden changes in surface temperatures over short distances along the east coast of continents can cause a concentration of solenoids which can easily generate a cyclonic curvature. The Hatteras low is a prime example of this type of cyclogenesis.

5. Winds aloft flowing parallel to fronts at the surface. When winds aloft up to about 700 mb flow parallel to fronts on the cold air side, vorticity considerations often lead to the development of waves along such fronts and the formation of new cyclones.

The vast majority of cyclones, it has just been pointed out, develop as a result of wave action along fronts. There are additional causes for cyclogenesis; they are as follows:

1. Barrier of cold air. (The barrier theory of cyclogenesis.) Cyclogenesis may occur when a slow moving mass of cold air blocks a rapidly eastward moving mass of warm air by dynamic pressure reduction to the "lee" of the cold air. The lee in this case is the return flow meeting the advancing warm high. This type of cyclogenesis is similar to that resulting from mountain waves.

2. Convection. (The convection theory of cyclogenesis.) The convection theory suggests that widespread intense convection may cause cyclogenesis as a result of the influx of air at the surface if the convection is of sufficient duration and intensity.

Barrier and convection cyclogenesis do not contribute greatly to the formation of new lows.

The last type of extratropical cyclogenesis to consider is the formation of thermal lows. These lows develop on a large scale over the continental subtropics in summer. The local heating lifts the isobaric surfaces in the air over the heat source. This results in an upper level high and an outflow of air aloft and inflow at the surface. These thermal lows, although they possess cyclonic flow in the low levels, are more akin to stationary anticyclones in their behavior and in their influence on the weather.

CYCLOLYSIS

When pressure at the center of a cyclone increases at a greater rate than the surrounding area, the intensity of the cyclonic circulation decreases. This process is called CYCLOLYSIS. It can occur even before occlusion begins, in the case of a stable wave, and always takes place in mature occlusions. In the case of frontal cyclones, frontolysis also occurs during cyclolysis. Since air mass contrasts are least near cyclone centers, frontolysis of occluded fronts proceeds from the center outward.

ENERGY OF CYCLONES

Cyclones transform a great deal of potential energy into kinetic energy in the atmosphere. Their greatest source of energy is the potential energy of horizontal mass distribution; that is, the energy of air masses of different temperature lying side by side. When the frontal slope is not an equilibrium slope, potential energy exists. This energy is converted to kinetic energy as the air masses try to adjust themselves to equilibrium. The solenoids are tilted and therefore accelerate the circulation both in the vertical and in the horizontal. This acceleration continues as long as the favorable solenoidal field exists. Cyclones thus increase in energy until they reach maximum intensity at occlusion.

Two other major sources of energy in developing cyclones vie for second place, depending upon the origin of the air masses, the moisture content, and their relative stability. They are the kinetic energy of the surrounding air and the energy from the latent heat of condensation.

The kinetic energy of the surrounding air is derived from the concentration of air over a smaller area as a result of horizontal convergence toward the center and a resultant increase in the wind speeds.

The latent heat of condensation is released when the moisture in the air condenses as a result of lifting. It is a large contributor in extratropical cyclones and the major source of energy in tropical cyclones.

TYPES OF WAVE CYCLONES

The change in wind velocity along a horizontal axis perpendicular to the direction of flow is called horizontal wind shear. The six possible combinations of vector pairs which involve shear are illustrated in figure 5-23.

The upper row shows the three pairs which give cyclonic shear, while the lower row shows the opposite or anticyclonic shear combinations. Needless to say, only the A, B, and C types are possible in frontal troughs.

There is some doubt as to whether type C wave cyclones actually occur, especially in middle

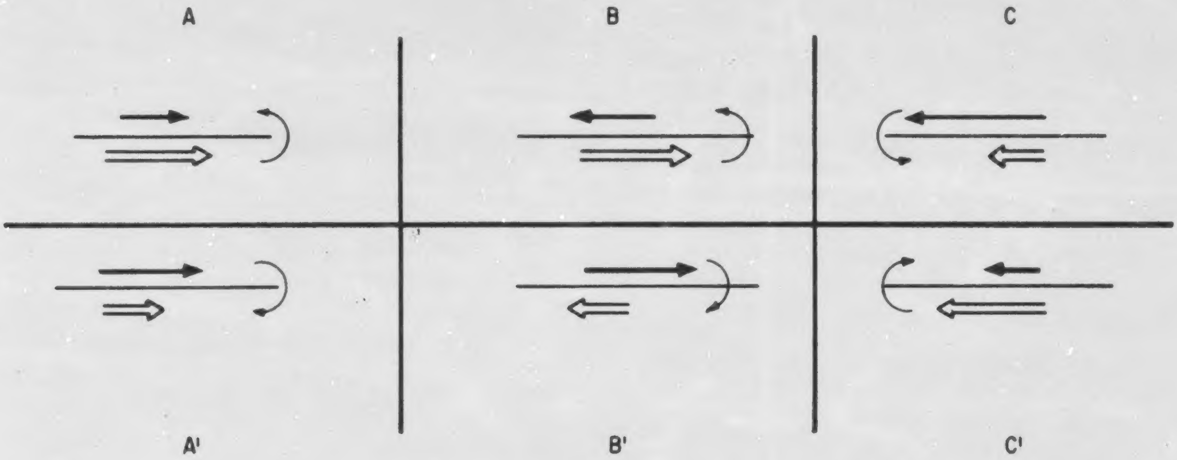

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Figure 5-23.–Six possible combinations of vector pairs involved in shear.

latitudes where synoptic networks are dense enough to make positive identification possible. The most likely areas of formation of type C wave cyclones would be the trailing ends of polar fronts in the subtropics and the Arctic front. Winter is therefore the only season when such disturbances could occur. Cyclones do develop out of type C shear, but they are nonfrontal. Tropical disturbances associated with waves in the easterlies are of this type. Type C wave cyclones, if they occur, would never occlude, since their life cycle must be short and their closed circulations, if any, are of limited extent.

The most common, called the classical wave cyclone, is type B. Its life cycle as illustrated in figure 5-21 shows the occlusion process. This cyclone often moves, deepens, and occludes with great rapidity. It is most likely to exhibit the classical frontal characteristics. Figure 5-24 shows an idealized model with cyclonically curved isobars in the cold air mass and straight parallel isobars in the warm sector.

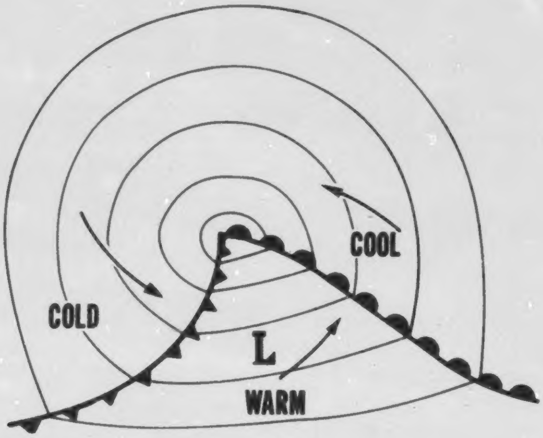

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Figure 5-24.—Idealized type B cyclone.

The warm front generally increases in intensity while the cold front becomes more diffused. The isobar kinks, especially along the cold front, are slight. The warm sector is quite homogeneous with respect to temperature. The most frequent modifications and exceptions are discussed later in the chapter.

Of less frequent occurrence, but fairly common, is the type A wave cyclone. This type cyclone is illustrated in figure 5-25(A).

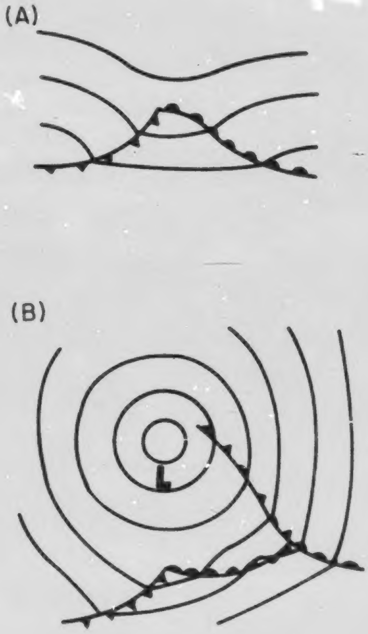

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Figure 5-25.—Type A wave cyclone. (A) Typical isobaric configuration; (B) typical appearance of type A wave cyclone in a synoptic situation.

It moves more rapidly than the type B wave but seldom deepens or occludes. Its cold air isobars are anticyclonically curved and its pressure field, as a whole, is symmetrical. Frontogenesis along the cold front and frontolysis along the warm front are common (note isobaric curvature). This type of wave cyclone normally occurs in the synoptic situation, shown in figure 5-25(B) as the second or third member of a frontal cyclone family. Rapid cyclogenesis often occurs when the wave overtakes the occluded system ahead of it.

OCCLUSION OF UNSTABLE WAVE CYCLONES

The Polar Front Theory and the resultant occlusion of unstable waves was discussed in a previous section of this chapter. The position, orientation, and season of the year have a great bearing on whether the cyclone will occlude or remain a stable wave.

The general orientation of the frontal system along which a wave cyclone moves usually determines whether or not it will occlude and if so, what type of occlusion is most likely to form, provided the underlying surface is uniform; that is, all ocean or all land.

In figure 5-26, the wave on the left is unlikely to occlude at all, the one in the center will become a cold type occlusion if it occludes, and the wave moving poleward is apt to become a warm type occlusion.

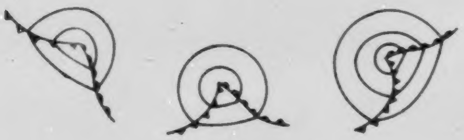

Figure 5-26.—Likelihood of occlusion in relation to orientation of the wave cyclone.

Type A waves, if occlusion occurs, would show very little contrast across the front, with slightly greater change of colder air to the westward. However, actual data may occasionally show conditions to the contrary and the actual observations should take precedence.

Occlusions are discussed in detail in chapter 6 of Aerographer's Mate 3 & 2 and some of the characteristics of occlusions are discussed later in this chapter.

RELATION OF CYCLONES TO UPPER AIR FEATURES

Figure 5-27 illustrates a typical model unoccluded wave cyclone in relation to the thickness lines, upper troughs, and upper contours.

The simple model of a warm sector frontal depression is given in this figure. The upper portion represents the 1,000-500 mb thickness field and the fields of the 1,000 and 500 mb contours. A pronounced warm tongue of thickness is associated with the warm sector depression, and the thickness lines are widely spaced over the warm sector. The thickness lines are nearly parallel to the sea level fronts, and their strongest gradients lie at some distance on the cold sides of the fronts. The 500-mb contours show a pronounced ridge just ahead of the sea level low center. A belt of maximum 500-mb geostrophic winds, which appears in the 500-mb contour pattern, coincides with the strongest concentration of 1,000-500 mb thickness lines; that is, it lies on the warm side of the frontal zone at 500 mb.

A vertical cross section of thickness along line X-Y is given in the middle portion of figure 5-27. For simplicity, the 1,000-mb surface is represented as a horizontal line. There is very little thickness gradient above sea level in the warm sector. Strong thickness gradients are observed between sea level and 500-mb locations of both cold and warm fronts.

The lower portion of figure 5-27 illustrates the height variations above sea level of the 1,000-, 700-, and 500-mb surfaces. The trough and ridge axes slope northwestward with height. A strong 500-mb contour gradient appears in the zone bounded by the positions of the front at 1,000 and 500 mb.

In general, frontal slopes are steepest, hence thickness gradients are greatest near the warm sector apex.

SECONDARY CYCLONES

The second and younger member of a family of cyclones on a polar front is called a secondary cyclone. According to our previous classifications of types A, B, and C, the first two were discussed in a previous section of this chapter and are classified as secondary cyclones. Another type of secondary cyclone was named by its Norwegian discoverers for the area where it was most likely to occur. This type is known as the Skagerrak. It was originally thought to have been caused by the orography of the southern Scandinavian peninsular. Similar occurrence at other orographically preferred areas of the world have further confirmed this theory, but it is known to occur even over the open ocean areas where some of the factors must be operative. Typical stages in the process are shown in figure 5-28.

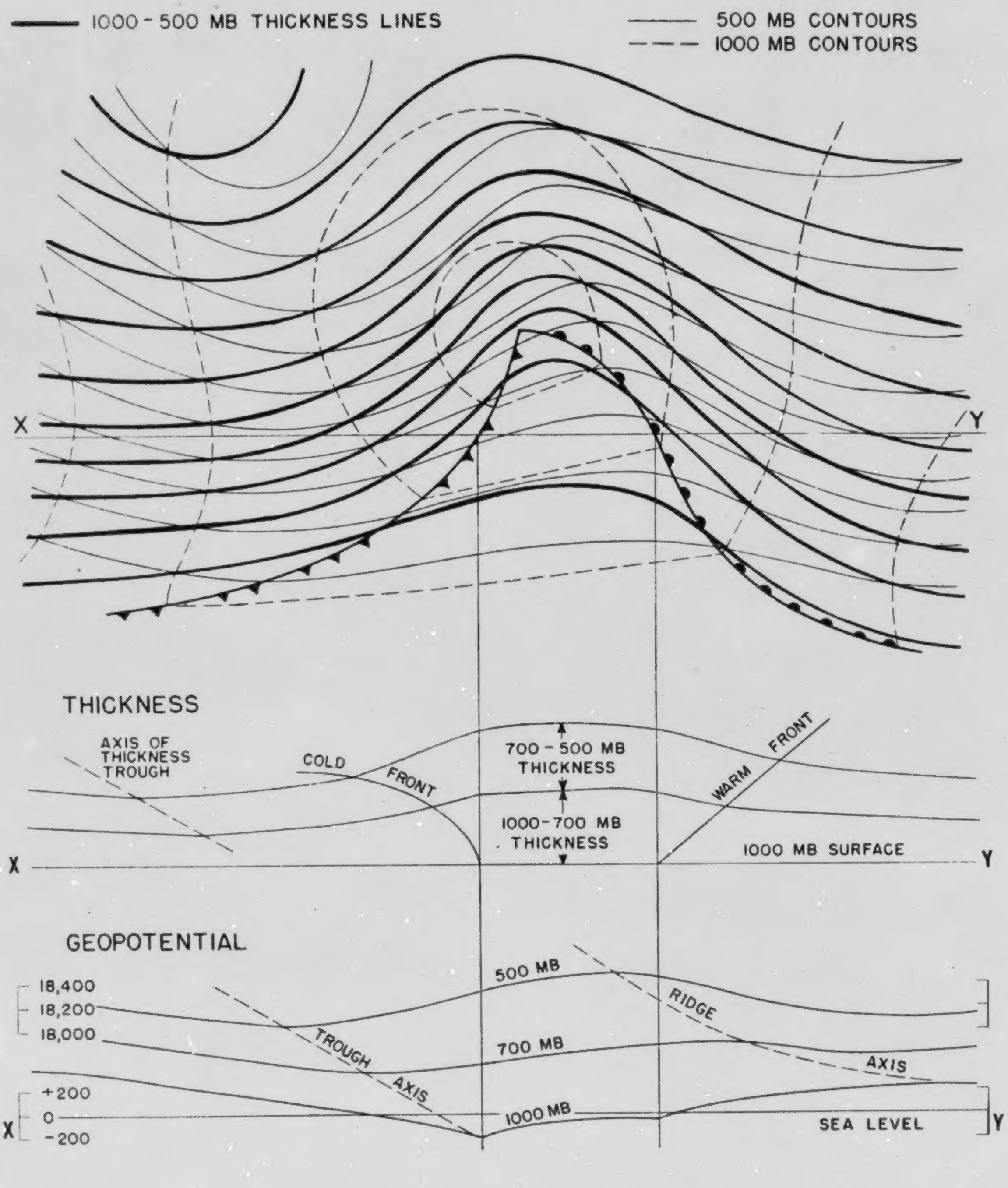
1000 - 500 MB THICKNESS LINES
500 MB CONTOURS
1000 MB CONTOURS
X
Y
THICKNESS
AXIS OF THICKNESS TROUGH
COLD FRONT
700 - 500 MB THICKNESS
1000-700 MB THICKNESS
WARM FRONT
1000 MB SURFACE
X
Y
GEOPOTENTIAL
18,400
18,200
18,000
500 MB
RIDGE
AXIS
TROUGH
AXIS
700 MB
+200
0
-200
1000 MB
SEA LEVEL
X
Y

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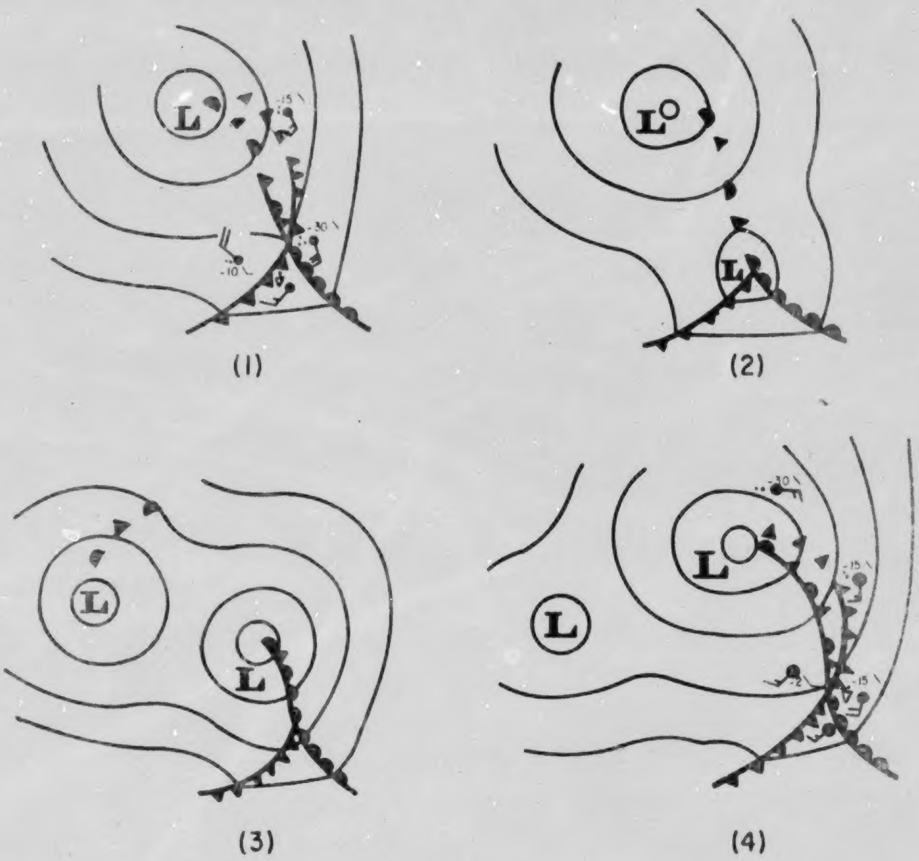
Figure 5-27.—A simplified 3-dimensional model of a wave cyclone showing relationship of upper air features.



(1) (2)

(3) (4)

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Figure 5-28.—Stages in the development of a Skagerraking cyclone.

Principal surface indications of this type of cyclogensis are:

1. Spreading of isobars north of the peak of a warm sector.
2. Large negative tendencies near the peak, especially in the warm air.
3. Diverging isobars in the warm sector.
4. Increasing intensity of precipitation near the peak.
5. Increasing cross-isobaric flow around the peak.

Orography, without a doubt, plays a great role in certain preferred areas of Skagerraking, but over the ocean some other factors must be operative. In some cases, a rapidly moving type A wave which overtakes the slow moving occlusion may be the triggering mechanism for cyclogenesis.

Whatever the exact nature of its causes, this type of cyclogenesis proceeds with great rapidity. The new occlusion forms immediately, and soon overshadows its predecessor in both area and intensity. However, the cold occlusion, having greater vertical extension, exerts a certain control on the movement of the new center, which at first follows the periphery of the old center. Later, the two centers pivot cyclonically about a point somewhere on the axis joining them until the old center has filled and loses its separate identity.

Skagerraking can take place with either the warm or cold type occlusions. If it occurs near a west coast in winter, there is a good chance the new occlusion will be of the warm type.

WARM SECTOR TROUGHS

One of the most common ways in which frontal cyclones differ from the idealized model is through the presence of a nonfrontal trough in the warm air. Two examples are illustrated in figure 5-29(A) and (B).

The front in (A) can be quite intense, but the cold front in (B) is generally weak and diffused since the prefrontal trough is more marked than the frontal trough. The trough in (A) can on occasion contain a cold front, in which case its extension north of the principal front should be carried as an upper cold front. Cyclones in low latitudes in particular, commonly have prefrontal troughs of the type shown in (B). Although most of the definite examples of this type of wave cyclone are found in the dense synoptic networks of populated areas, there is no reason to believe that they occur only over land. Here, close attention to indications of a prefrontal trough must be given when making an oceanic analysis.

SUMMARY

Table 5-1 presents average values of certain numerical characteristics of various stages in the life cycle of an unstable wave cyclone.

Data given are to be used only as a general guide and principally in areas where reports are sparse. Actual conditions may deviate greatly from these averages. This table applies only to wave cyclones which are occluding. Nonfrontal cyclones Skagerraks behave in an entirely different fashion.

The usefulness of this table is evidenced by the fact that if one or two of the listed characteristics are known, then the table may be used to fill in the mission features.

For an example of the use of this table, assume that a low-pressure center has been moving in a northeast direction at about 25 knots for the past 18 hours. The table shows that it should be considered an occlusion with the central pressure somewhat under 1,000 mb.

THE COLD FRONT

Cold fronts usually move faster and have a steeper slope than warm fronts. Cold fronts which move very rapidly have very steep slopes in the lower levels and narrow bands of clouds which are predominant along or just ahead of the front. Slower moving cold fronts have less steep slopes and their cloud systems may extend far to the rear of the surface position of the fronts. Both fast moving and slow moving cold fronts may be associated with either stability or instability and either moist or dry air masses. The typical characteristics of the fast and slow

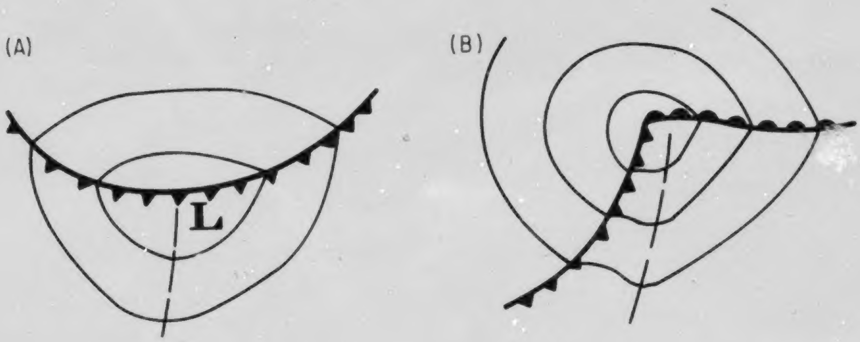

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Figure 5-29.—Two examples of a trough in the warm sector. (A) Cold front oriented E-W; (B) cold front oriented NNE-SSW.

Table 5-1.–Numerical characteristics of the life cycle of an unstable wave cyclone.

	Wave Cyclone	Occlusion	Mature Occlusion	Cyclolysis
Time (hours)	0	12-24	24-36	36-72
Central Pressure (mb)	1,012-1,000	1000-988	984-968	998-1,004
Direction of Movement (toward)	NE to SE or (quad)	NNE to N (arc)	N to NNW (arc)	↺
Speed of Movement (knots)	30-35	20-25	10-15	0-5

The ↺ symbol indicates that the filling center drifts slowly in a counterclockwise direction along an approximately circular path about a fixed point.

moving cold fronts are discussed in detail in chapter 6 of AG 3 & 2. We are concerned here chiefly with the upper air characteristics of these fronts.

Both slow and fast moving cold fronts develop when cold air bulges southward along the frontal boundary. The inclination of the frontal slope depends chiefly upon the wind direction and velocity arrangement across the front.

SLOW MOVING COLD FRONT

The slope of this type front is usually in the neighborhood of 1:100 miles. Near the ground the slope is often much steeper due to surface friction.

The type of weather experienced with this front is dependent upon the stability of the warm air mass. When the warm air mass is stable, a rather broad zone of altostratus and nimbostratus cloud systems follow the front by several hundred miles. If the warm air is unstable (or conditionally unstable), thunderstorms and cumulonimbus clouds may develop in this cloud bank and may stretch for some 50 miles behind the surface front. These clouds all form within the warm air mass. In the cold air there may be some stratus or nimbostratus formed by the falling rain but generally outside the rain areas there are relatively few low clouds. This is due to the descending motion of the cold air which produces, at times, a subsidence inversion some distance behind the front.

The type of precipitation observed is also dependent upon the stability and moisture conditions of the air masses.

The ceiling is generally low with the frontal passage, and gradual lifting is observed after passage. Visibility is poor in precipitation and may continue low for many hours after frontal passage as long as the precipitation occurs. When the cold air behind the front is moist and stable, a deck of stratus clouds and/or fog may persist for a number of hours after frontal passage.

Characteristic Upper Air Features

TEMPERATURE.–The temperature inversion on this type front is usually well marked. In the precipitation area the relative humidity is high in both air masses. Farther back of the front, subsidence may occur, giving a second inversion closer to the ground.

Upper air contours are usually parallel to the front as well as the mean temperature (thickness lines). The weather will usually extend as far in back of the front as these features are parallel to it. When the orientation changes, this usually indicates the postion of the upper air trough.

WINDS.–The wind usually backs rapidly with height (on the order of some 60 to 70 degrees between 950 and 400 mb), and at 500 mb the wind direction is inclined at about 15 degrees to the front. The wind component normal to the front decreases slightly with height, and the component parallel to the front increases rapidly. The thermal wind between 950 and 400 mb is almost parallel to the front.

Surface Characteristics

The pressure tendency associated with this type frontal passage is usually indicated by either an unsteady or steady fall prior to frontal passage, while the rises behind this type front are weak. Temperature and dewpoint drop sharply with the passage of a slow moving cold front. The wind veers with the cold frontal passage, reaches its highest speed at the time of frontal passage. Isobars may be curved anticyclonically in the cold air. This type front usually moves at an average speed between 10 and 15 knots.

Figure 5-30 illustrates the typical characteristics in the vertical of a slow moving cold front (upper half) and typical upper air flow in back of the front and accompanying surface weather (lower half).

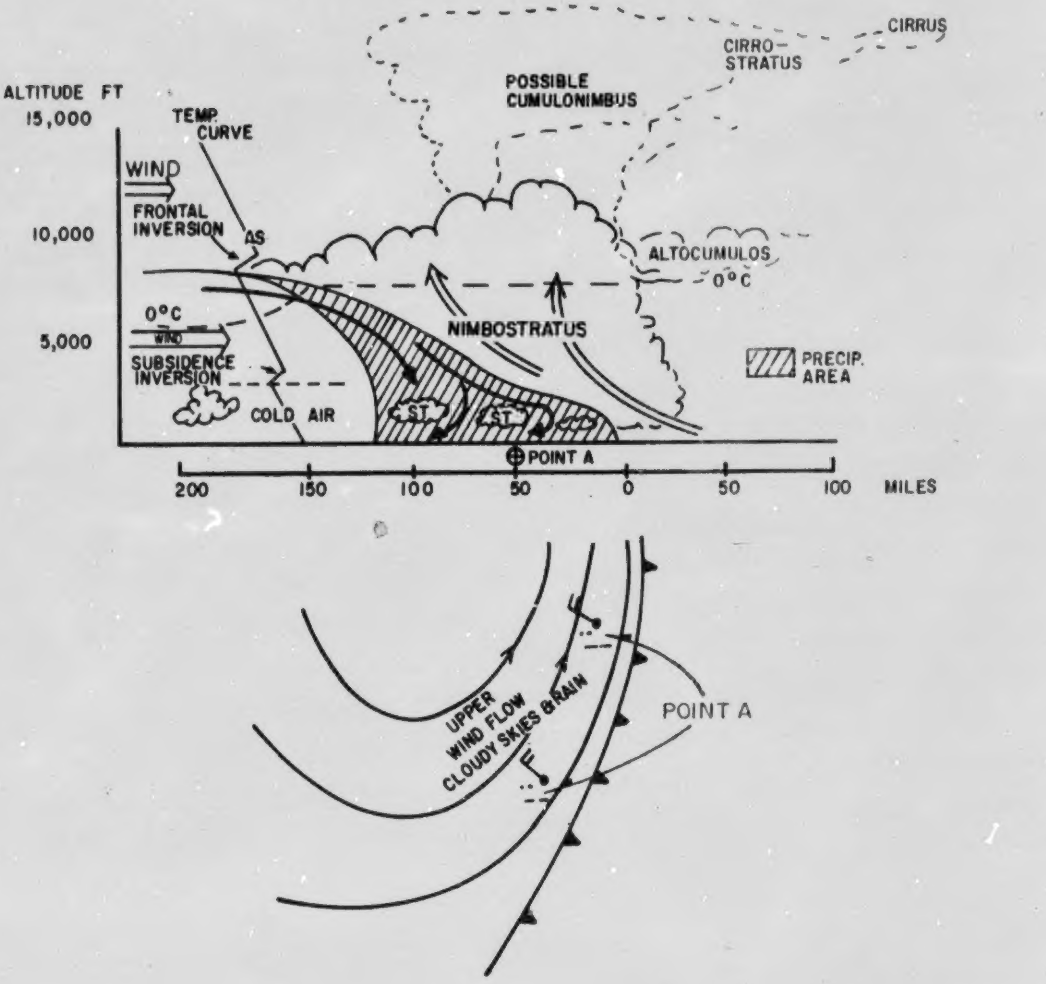
CIRRUS
CIRRO-STRATUS
POSSIBLE CUMULONIMBUS
ALTITUDE FT
15,000
TEMP. CURVE
WIND
FRONTAL INVERSION
AS
10,000
ALTOCUMULOS
0°C
0°C
NIMBOSTRATUS
5,000
WIND
SUBSIDENCE INVERSION
PRECIP. AREA
ST
ST
COLD AIR
POINT A
200
150
100
50
0
50
100
MILES
UPPER WIND FLOW
CLOUDY SKIES & RAIN
POINT A

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Figure 5-30.—Typical vertical structure of a slow moving cold front with upper windflow in back of the front.

Since this is only one typical case, many variations to this model can occur in nature.

FAST MOVING COLD FRONT

The warm air above the frontal slope is moving faster than the cold air. This type front is characterized by a narrow band of weather associated with the downslope motion of the warm air above the frontal slope. The descending air is warmed adiabatically and the temperature contrast across the front is increased, thus strengthening the frontal discontinuity. This front has an average slope of 1:40 to 1:80 and moves with an average speed of about 25 knots. This is the most important type of cold front.

If the warm air is moist and unstable, a line of thunderstorms frequently develops along this type front. Sometimes, under these conditions, a line of strong convective activity is projected 50 to 200 miles ahead of the front and parallel to it. This may develop into a line of thunderstorms called a squall line. On the other hand when the warm air is stable, an overcast of altostratus clouds with general rain may extend over a large area ahead of the front. If the warm air is very dry, little or no cloudiness is associated with the front. The front depicted in the following sections is a typical front with typical characteristics.

Weather

Cumulonimbus clouds are observed along and just ahead of the surface front. Stratus, nimbostratus, and altostratus may extend ahead of the front from the cumulonimbus. These clouds may extend as much as 150 miles ahead of the front, but they are generally the altocumulus and stratocumulus type. The clouds just mentioned are all in the warm air. Generally, unless the cold air is unstable and descending currents are weak, there is a lack of clouds in the cold air behind the front. Showers and thundershowers occur along and just ahead of the front. The ceiling is low only in the vicinity of the front. Visibility is poor in precipitation but improves rapidly after passage.

Upper Air Features

TEMPERATURE.–Owing to the sinking motion of the cold air behind the front and the resultant adiabatic warming, the temperature change across the front is often destroyed or may even be reversed. A sounding taken in the cold air immediately behind the surface front would indicate only one inversion with the characteristic inversion and an increase in moisture through the inversion. Farther back of the front, a double inversion structure would be in evidence. The lower inversion would be due to the subsidence effects in the cold air. This is ofttimes confusing to the analyst as the subsidence inversion is usually more marked than the frontal inversion and may be mistaken for the frontal inversion.

WINDS.–In contrast to the slow moving cold front, wind above the fast moving cold front exhibits only a slight backing with height on the order of some 20 degrees between 950 and 400 mb and the wind direction is inclined toward the front at an average angle of about 45 degrees. The wind components normal and parallel to the front increase with height, and the wind component normal to the front exceeds the mean speed of the front at all levels above the lowest layers. The thermal wind for the 950-400 mb layer has an average angle of about 30 degrees to the front.

Surface Characteristics

Pressure tendency falls ahead of the front, and there are sudden and strong rises after the frontal passage. If a squall line lies some distance ahead of the front, there may be a strong rise associated with its passage and a shift in the wind. However, after the influence of the squall line has passed, winds will back to southerly and pressures will level off. The temperature shows a fall in the warm air just ahead of the front due to evaporation of falling precipitation. Rapid clearing and adiabatic warming just behind the front tend to keep the cold air temperature near that of the warm air. An abrupt temperature change usually occurs far behind the front. The dewpoint and wind direction are better indicators of the passage of a fast moving cold front. The wind veers with frontal passage and is strong, gusty, and turbulent for a considerable period of time after passage. Dewpoint also decreases sharply after passage.

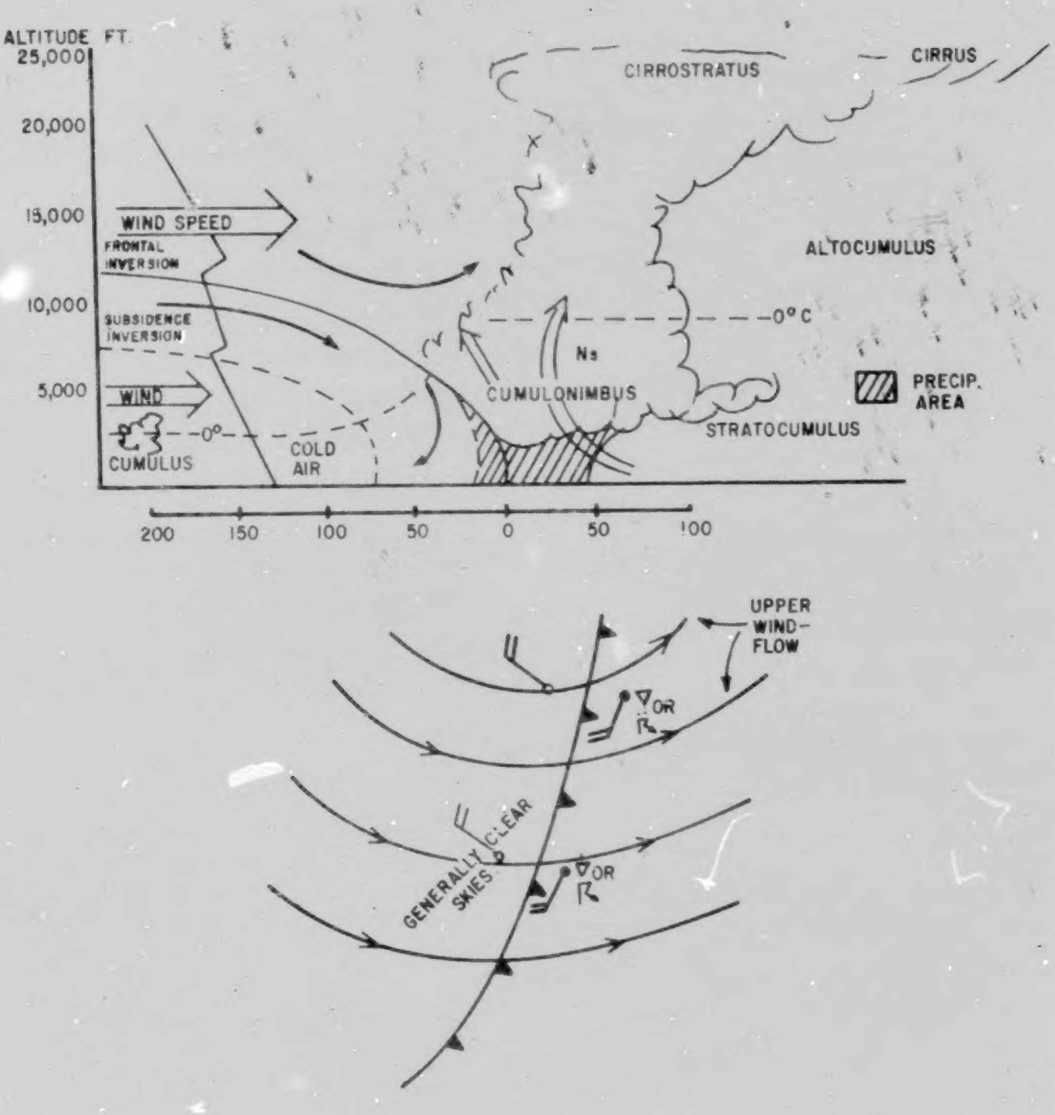
ALTITUDE FT.
25,000
20,000
15,000
10,000
5,000
WIND SPEED
FRONTAL INVERSION
SUBSIDENCE INVERSION
WIND
0°
CUMULUS
COLD AIR
CIRROSTRATUS
CIRRUS
ALTOCUMULUS
0° C
Ns
CUMULONIMBUS
STRATOCUMULUS
PRECIP. AREA
200
150
100
50
0
50
100
UPPER WIND-FLOW
OR
GENERALLY CLEAR SKIES
OR
AG.485

Figure 5-31.—Typical vertical structure of a fast moving cold front with upper windflow across the front.

Figure 5-31 illustrates a vertical cross section of a fast moving cold front with resultant weather. Also indicated in the lower half of the diagram is the surface weather at a point a short distance in advance of the front and the upper airflow above the front.

OTHER UPPER AIR CHARACTERISTICS OF COLD FRONTS

Cold fronts on upper air charts are characterized by a packing of isotherms behind them. The more closely packed the isotherms, and the

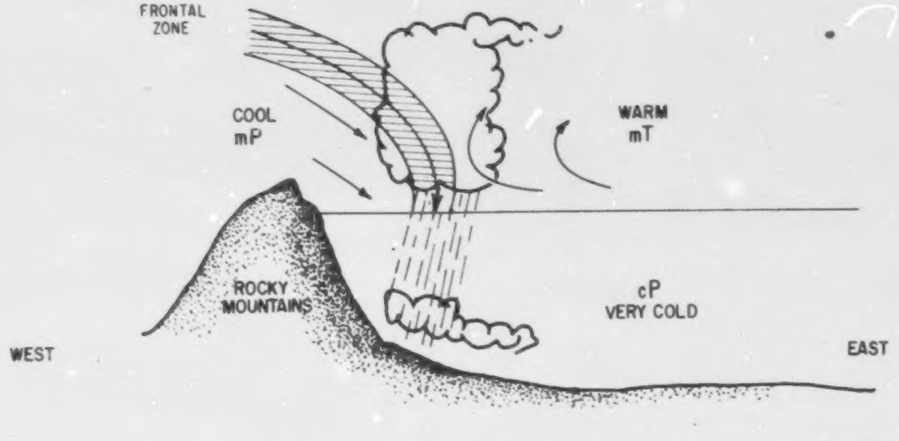

Figure 5-32.—Upper cold front.

more nearly they parallel the fronts, the stronger the front. When isotherms cross a front, they kink toward the cold air.

It is more common for a cold front on the surface to lie ahead of the upper air trough. This is illustrated in figure 5-31.

UPPER COLD FRONTS

It is generally recognized that there are two types of upper cold fronts. One is the upper cold front associated with the warm type occlusion which will be taken up later in this chapter. The other occurs most frequently in the area just east of the Rocky Mountains in winter when mP air crosses the mountains behind a cold front or behind a trough aloft. Usually a very cold layer of continental polar air is lying next to the ground over the area east of the mountains. Warm maritime tropical air moving northward from the Gulf of Mexico has been forced aloft by the cold cP air. When the cool mP air flows over the mountains, it forces its way under the warm mT air aloft and flows across the upper surface of the cP air just as if it were the surface of the ground. All frontal activity in this case takes place above the top of the cP layer. Refer to figure 5-32 for an example of this type of front.

Weather from this type of front can produce extensive cloud decks and blizzard conditions for several hundred miles over the midwestern plains.

SQUALL LINES AND INSTABILITY LINES

A squall line is a line of active showers and thundershowers that roughly parallels the cold front and generally occurs in the warm sector of wave cyclones. This type usually develops along or ahead of fast moving cold fronts. The term squall line is often confused with the term instability line. Although the two are often used interchangeably, their meaning has been defined by agreement of the Subcommittee on Aviation Meteorology of the Air Coordinating Committee. Their definitions are as follows:

1. INSTABILITY LINE. This is a line of incipient, active, or dissipating nonfrontal instability conditions; it is an analytical term for indicating primarily the incipient and dissipating stages of nonfrontal squall line phenomena and for the sake of continuity also includes the active squall line stage. It is frequently found in the warm sector of an extratropical cyclone, and unlike a true front the instability line is transitory in character, usually developing to maximum intensity within a period of 12 hours or less and then dissipating in about the same length of time.

2. SQUALL LINE. A squall line is a line of active thunderstorms or squalls which may extend over several hundred miles. It is the phenomenon of the mature and active stage of instability line development and may be either a solid or broken line of numerous thunderstorms accompanying vertical motions of a greater order of magnitude than is usually present in the atmosphere.

NOTE: The term instability line, as defined above, is the more general term and includes squall line as a special case. The two will be used interchangeably in this section of the chapter.

In this section general characteristics of squall lines occurring in the warm sector in advance of fast moving cold fronts and the more special case of the instability line not associated with a fast moving cold front over the Plains States are discussed.

Squall lines, develop ahead of fast moving cold fronts in the warm sector of a wave cyclone. This line is located, on the average, some 100 to 300 miles ahead of the cold front.

A typical case showing the isobaric patterns is shown in figure 5-33.

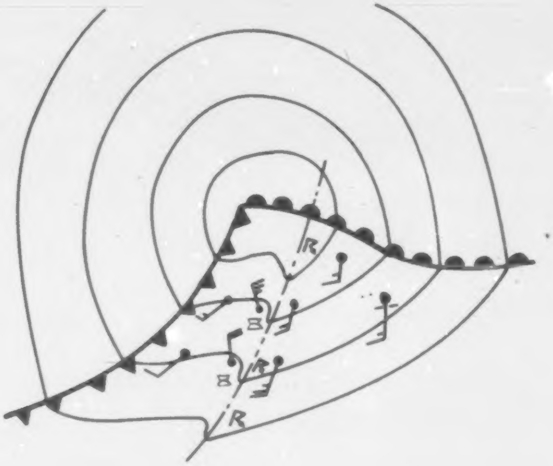

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Figure 5-33.—Typical isobaric pattern associated with a warm sector squall line.

Showers and thunderstorms (sometimes tornadoes) occur along the squall line, and the wind shifts cyclonically with passage. However, if the zone is narrow, the wind shift may not be noticeable on surface charts. There is generally a large drop in temperature due to the cooling of the air by precipitation. Pressure rises after the passage of the squall line, and at times a small micro-high may form behind it. The cold front has little weather or clouds associated with it. After passage of the squall line, the wind will back to southerly before the cold frontal passage.

Squall lines move along in advance of the cold front at a speed at times exceeding that of the associated cold front. A rough guide to its direction and speed can be obtained from the 500-mb winds. It will move approximately 40 percent of the wind speed and along the 500-mb winds.

Vertical Structure

Theoretical explanation of the phenomenon is still incomplete, but since it is known to be associated with fast moving cold fronts, the winds at upper levels are moving faster than the winds in the cold air beneath the frontal surface. A vertical cross section of the squall line in the warm sector is depicted in figure 5-34.

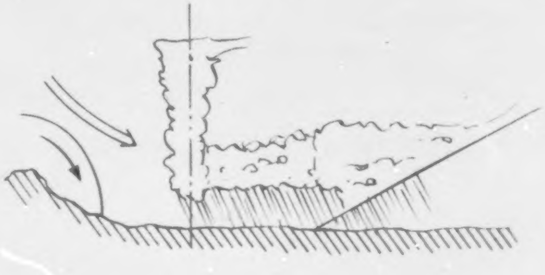

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Figure 5-34.—Vertical cross section of squall line associated with a fast moving cold front.

A typical phenomenon noted in this illustration shows to the east of the squall line pseudo-warm frontal weather.

It should be noted that weather associated with the cold front decreases in intensity after a squall line has formed and increases in intensity after the squall line has dissipated.

The squall line is not necessarily confined to the warm sectors of wave cyclones. In some instances it has been first identified behind the cold front which it overtakes and passes as it moves out in the warm sector. It can also extend poleward of the warm front, and often retains its identity during the occlusion process. This can lead to the curious situation of a cold occlusion with an apparent upper cold front, the squall line.

Development

The following conditions are favorable for squall line development of this type:

1. The warm sector, or other area of suspicion, is watched for development of a trough or line of cyclonic wind shear and for actual development of thunderstorms.
2. Cold air advection in the middle and higher levels.
3. Minimum stability indexes are usually zero or less in the area of squall line formation. However, the line may form east or north and downwind some 100 to 300 miles from this area.
4. Maximum dewpoint areas at the 850- and 700-mb levels should be noted because squall lines form in these areas (when other conditions are favorable) some 90 percent of the time.
5. A favored position for squall line formation is on the west and central portion of moist and warm tongues.
6. Increasing moisture at low and high levels with a midlevel dry source upwind. This permits evaporative cooling.
7. Low-level convergence and lifting.
8. The most favorable geographical area in North America is the Central and Southern States in late winter and spring. (Farther northward the season runs from late spring through summer.)
9. Warm air advection at 850 mb.
10. In the cyclonic shear of the jet at 850 mb.
11. A mechanism to divert the strong middle level winds to the surface. This mechanism is believed to produce an autoconvective lapse rate by rapid evaporation of moisture in a mixing zone. It usually consists of a moisture ridge oriented at a large angle to the windflow with a wind component across this ridge of at least 30 knots. The second mechanism is a source of dry air which the middle level wind carries to a position where it can be cooled by precipitation from above to its wet bulb temperature which must be lower than that of the moist air.
12. Strongly curved 200-300-mb flows are unfavorable for squall line development. A straight flow or slightly anticyclonic flow is more favorable.

Squall Line

Figure 5-35 illustrates a generalized, vertical cross section of the formation of the formidable Great Plains type squall or instability line not associated with a fast moving cold front.

The basic requirement for this type of formation is evaporative cooling of dry air as a result of high altitude precipitation.

An additional requirement is that surface heating be of sufficient intensity to cause thermal currents to transport moisture from lower to higher levels.

Formation of this type squall or instability line requires the following conditions to be fulfilled:

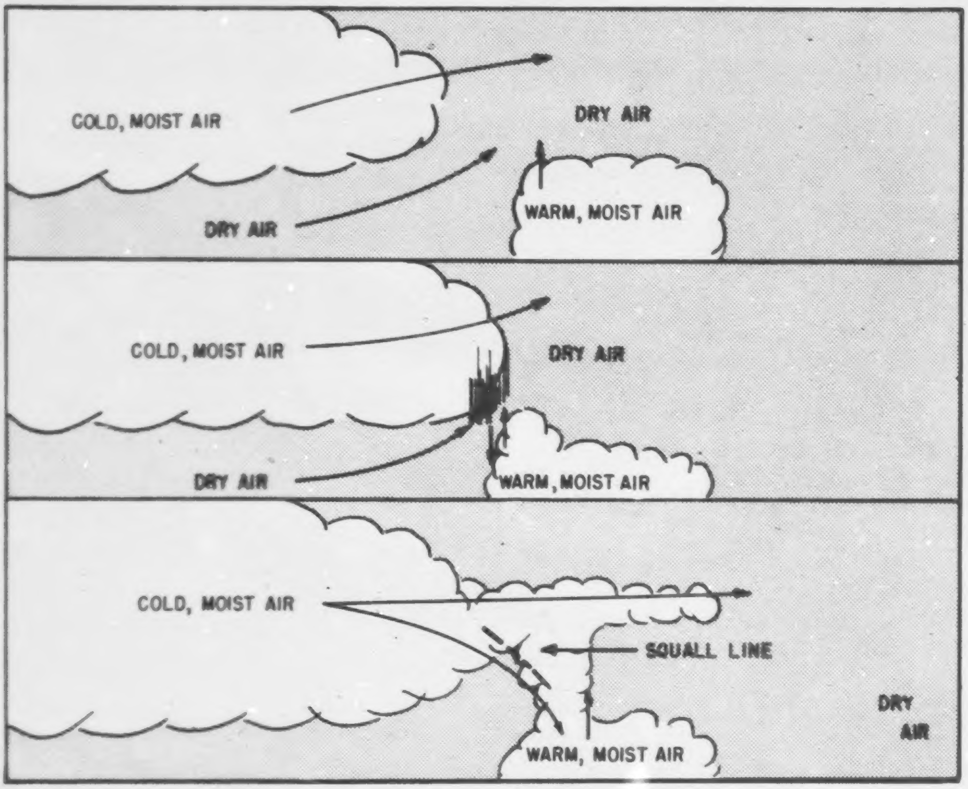
1. Cold moist air advection, usually at 14,000 feet msl or higher but based lower than 20,000 feet (top of fig. 5-35).
2. Cooling of the layers below the advection levels by evaporation of rain or cloud particles (middle of fig. 5-35).
3. An increase of surface temperature, usually by insolation, to the point that will cause convective currents to reach the condensation level.
4. Low-level wind convergence to increase updrafts (bottom of fig. 5-35).

WARM FRONTS

In this section some of the characteristic features of the warm front both at the surface and aloft will be discussed.

CHARACTERISTICS OF WARM FRONTS

The characteristics of a warm front will depend upon a number of factors. One is the



AG.489

Figure 5-35.—Formation of Great Plains squall line (not associated with fast moving cold front).

condition which existed prior to the lifting of the warm air; the stability of the air mass will determine the type of cloudiness that will form. The moisture content and the speed of movement of the two air masses are also determining factors as to the type and severity of the weather.

Slope

The slope of the warm front is usually somewhere between 1:100 and 1:300, with occasional fronts with lesser slopes. Therefore, warm fronts have characteristically shallow slopes which are due to the effect of surface friction which retards the frontal movement near the ground. There is a gradual veering of the wind through the frontal zone.

Movement

Warm fronts move slower than cold fronts. Their average speed is usually between 10 and 20 knots. The rule for determining the movement of warm fronts states that a warm front will move with a speed of 60 to 80 percent of the component of the geostropic wind normal to the front in the warm air mass.

Weather

The weather associated with typical warm fronts is given in chapter 6, AG 3 & 2.

The amount and type of clouds and precipitation vary with the characteristics of the air masses involved. Three situations are described in the following paragraphs:

1. When the overrunning warm air is moist and stable, nimbostratus clouds with continuous light to moderate precipitation are found approximately 300 miles ahead of the front. The bases of the clouds lower rapidly as additional clouds form in the cold air under the frontal surface. These clouds are due to evaporation of the falling rain. These clouds are stratiform when the cold mass is stable and stratocumulus when the cold air is unstable.

2. When the overrunning air is moist and unstable, cumulus and cumulonimbus clouds (thunderstorms) are frequently imbedded in the nimbostratus and altostratus clouds. In such cases, heavy rain showers, intense and intermittent, occur along with continuous precipitation.

3. When the overrunning warm air is dry, it must ascend to relatively high altitudes before condensation can occur. In these cases usually only high and middle clouds are observed.

Visibility is usually good under the cirrus and altostratus clouds. It decreases rapidly in the precipitation area. In addition, when the cold air is stable and extensive, fog areas may develop ahead of the front, and visibility is extremely low in this area.

Surface Features

The troughs associated with warm fronts are not as pronounced as those with cold fronts and sometimes make location difficult on the surface chart. The pressure tendency ahead of the front is usually a rapid or unsteady fall with a leveling off after frontal passage. A marked decrease in isallobaric gradient is noticed in the warm sector except when rapid deepening is taking place. The wind increases in velocity in advance of warm fronts due to an increase in pressure gradient and reaches a maximum just prior to frontal passage. The wind veers with frontal passage, usually from a southeasterly direction to a southwesterly direction behind the front.

Temperature generally is constant or slowly rising in advance of the front until the surface front passes, at which time there is a marked rise. This rise is dependent upon the contrast between the air masses. Dewpoint will usually increase slowly with the approach of the front with a rapid increase in precipitation and fog areas. If the warm sector air is maritime tropical, the dewpoint will show a further increase.

UPPER AIR CHARACTERISTICS OF WARM FRONTS

Upper Air Soundings

Warm fronts are not usually as well defined as cold fronts on upper air soundings. When the front is strong and little mixing has occurred, the front may show a well-marked inversion aloft, such as indicated in figure 5-36(A).

However, mixing usually occurs and the front may appear as a rather broad zone with only a slight change in temperature. Quite frequently there may be two inversions—one due to the front and the other due to turbulence. Lower cloud decks formed under the turbulence deck may obscure the altostratus frontal deck. (See fig. 5-36(B).)

Upper Air Charts

Isotherms are parallel to the front and show packing ahead of the front. The stronger the packing, the more active the front. The packing is not as pronounced as with the cold front.

A warm front is normally located in a slight trough along the edge of a ridge. This anticyclonic curvature aloft is due to vertical convergence, hence horizontal divergence occurs above the frontal surface.

UPPER WARM FRONTS

Upper warm fronts aloft are seldom encountered, but generally follow the same principles as cold fronts aloft. Once case when they occur is when the very cold air underneath a warm front is resistant to displacement and may force the warm air to move over a thinning wedge with a wave forming on the upper surface. This gives the effect of secondary upper warm fronts and may cause parallel bands of precipitation at unusual distances ahead of the surface warm front. When a warm front crosses a mountain range, it may encounter colder air to the east and may move along as a warm front aloft above the layer of cold air. This is common when a warm front crosses the Appalachian

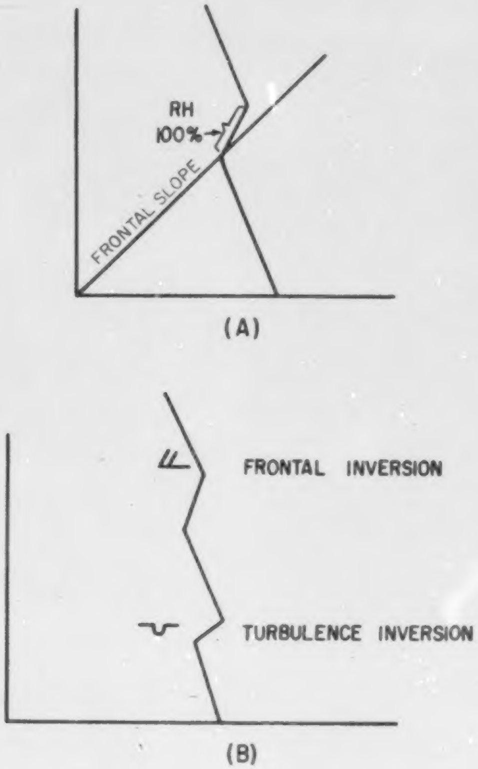


Figure 5-36.—Typical upper air soundings associated with a well-marked warm front. (A) Well-marked inversion; (B) frontal and turbulence inversions.

Mountains in winter. Warm air advection is more rapid and precipitation is heaviest where the steeper slope is encountered. Pressure falls rapidly in advance of the upper warm front and levels off underneath the horizontal portion of the front.

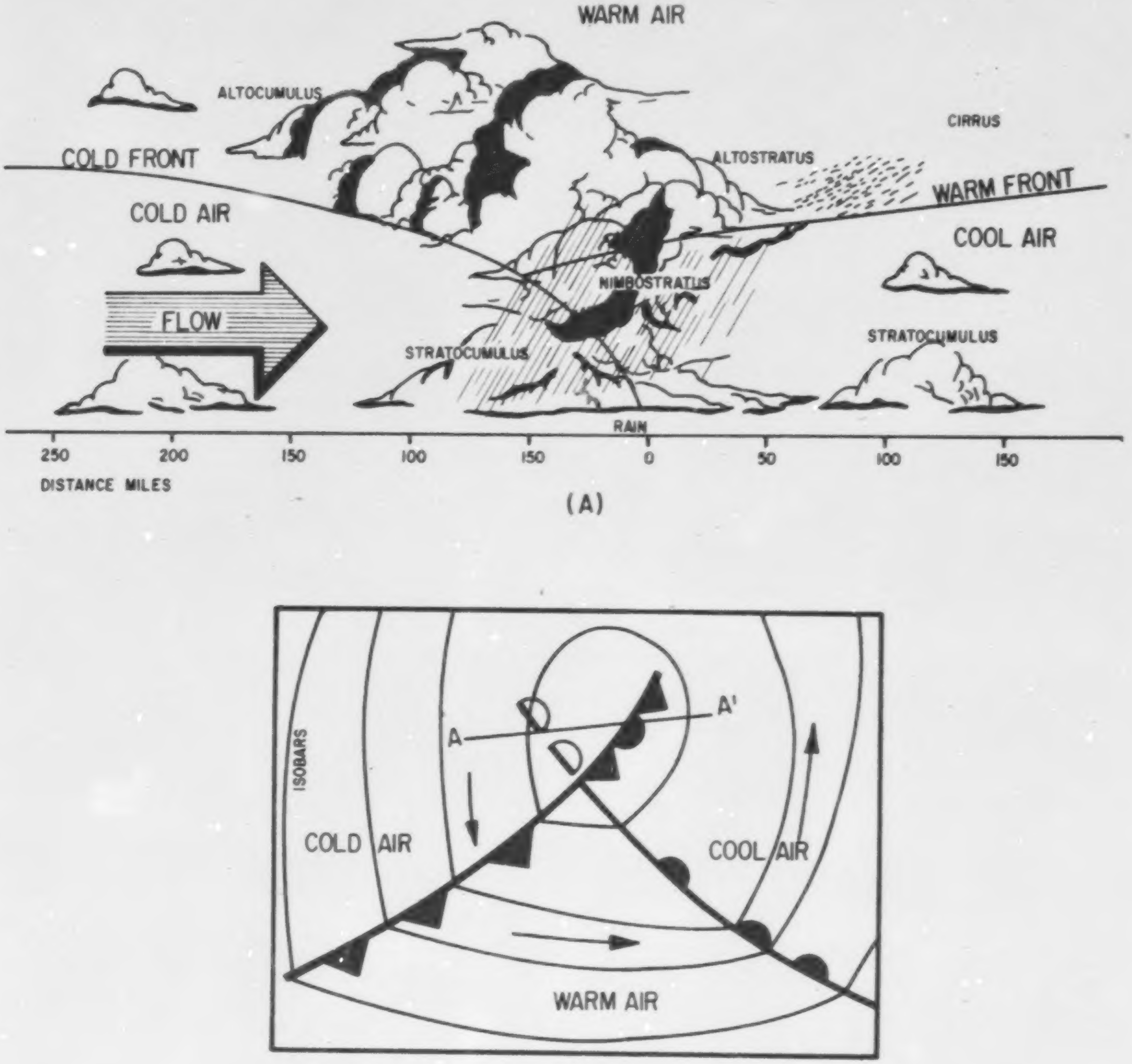
OCCLUDED FRONTS

The structure of the occlusion depends upon the temperature difference between the cold air in advance of the system and the cold air to the rear of the system. If the air in advance of the warm front is colder than the air to the rear of the cold front, when the cold air of the cold front overtakes the warm front, it will move up over this cold air in the form of an upper cold front, thereby forming a warm front type occlusion. When the air behind the cold front is colder, it will push in under the cool air in advance of the warm front and produce a cold front type occlusion. Vertical cross sections of these two types of occlusions are contained in chapter 6, AG 3 & 2.

COLD TYPE OCCLUSIONS

Formation

Cold front type occlusions form when the air ahead of the warm front is less cold than the air behind the overtaking cold front. When the cold front overtakes the warm front, both the warm air behind the warm front and the cool air ahead are lifted by the colder air moving in behind the cold front. Thus, the warm front itself is lifted


WARM AIR
ALTOCUMULUS
CIRRUS
COLD FRONT
ALTOSTRATUS
WARM FRONT
COLD AIR
COOL AIR
NIMBOSTRATUS
FLOW
STRATOCUMULUS
STRATOCUMULUS
RAIN
250
200
150
100
150
0
50
100
150
DISTANCE MILES
(A)
ISOBARS
A
A'
COLD AIR
COOL AIR
WARM AIR
(B)

AG.491

Figure 5-37.—Cold front type occlusion. (A) Vertical sturcture; (B) horizontal structure.

by the undercutting cold front, and it becomes an upper warm front. This front is seldom delineated on surface charts due to its close proximity to the surface front.

VERTICAL AND HORIZONTAL STRUCTURE.—Figure 5-37(A) illustrates a vertical cross section through the points A-A¹ on figure 5-37(B). Figure 5-37(A) shows the cold fron type occlusion in its initial stages when the cold air has begun to underrun the warm air. Figure 5-37(B) shows how this occlusion would be depicted horizontally on a surface chart.

WEATHER PHENOMENA.—In the occlusions initial stages of development, the weather and cloud sequence ahead of the occlusion is quite similar to that associated with warm fronts, while the cloud and weather sequence near the surface position of the front is similar to that associated with cold fronts. As the occlusion develops and the warm air is lifted to higher and higher altitudes, the warm front and prefrontal cloud systems disappear, and the weather and cloud system become similar to those of a cold front. Most of the precipitation occurs just ahead of the occlusion. Clearing behind the occlusion is usually rapid, especially if the occlusion is in the advanced stage. Otherwise, clearing may not occur until after the passage of the warm front aloft.

SURFACE INDICATIONS. The occlusion lies in a low-pressure area, and in the latter stages a separate low center may form at the tip of the occlusion, leaving another low-pressure cell near the end of the occlusion (see previous section on Skagerraking for an example). In this case the occlusion dissipates rapidly. The pressure tendency across the cold type occluded front follows closely those outlined for cold fronts, that is a leveling off, or more often a rapid rise after the passage of the occluded front.

Figure 5-38 shows the idealized frontal structure, isobaric and isallobaric patterns, and wind and weather distribution around the cold type occlusion.

Upper Air Characteristics

TEMPERATURE CURVES. If only one upper air sounding was taken so that it intersected either the cold or warm front, the sounding would appear as a typical warm or cold front sounding. However, if the sounding was taken so that it intersected both the cold and warm air, it would show two inversions. Figure 5-39 shows the typical lapse rate changes in a cold front type occlusion at the positions indicated.

In the three curves for station X, in figure 5-39(A), station X is successively in the cool air overrun by the warm air (curve A), in the warm sector (curve B), and in the cold air with subsidence (curve C). Station Y is deep within the circulation of the low, consequently the frontal discontinuities are weak due to convergent lift of the air. Successive positions of station Y, in A, the cool air is overrun by the warm air (curve A); in B, the cold air is overrun by the cool air (curve B); and in C, the station is in the cold air and reflects the convergent lift of air in that it has a slope near the moist adiabatic rate.

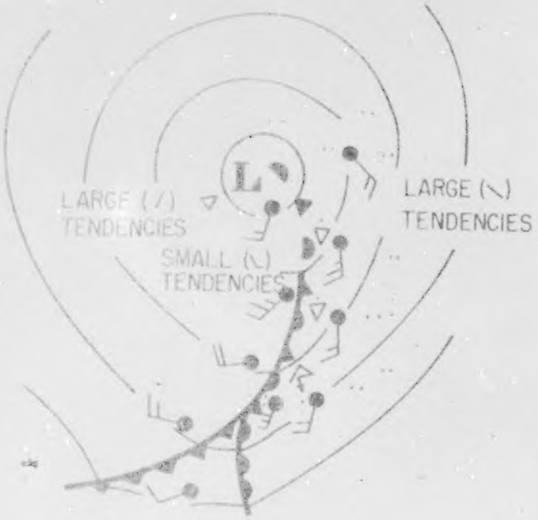


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Figure 5-38.—Typical cold type occlusion on the surface map.

UPPER AIR CHARTS. The occlusion may appear on some upper air charts. It will probably appear on the 850-mb chart, but rarely on the 700-mb chart. As the two air masses are brought closer together and as the occlusion process brings about gradual disappearance of the warm sector, the thickness gradient (mean isotherms) associated with the surface front weakens. The degree of weakening depends on the horizontal temperature differences between the cold air to the rear of the cold front and that ahead of the warm front. The angle at which the thickness lines cross the sea level position of the occluded fronts becomes greater as the temperature contrast between the two cold air masses decreases. A typical illustration of thickness lines (1,000-500 mb) would show a packing of isotherms in the cold mass behind the cold front, a lesser packing in the cool mass in advance of the warm front, and a warm ridge preceding the surface occlusion.

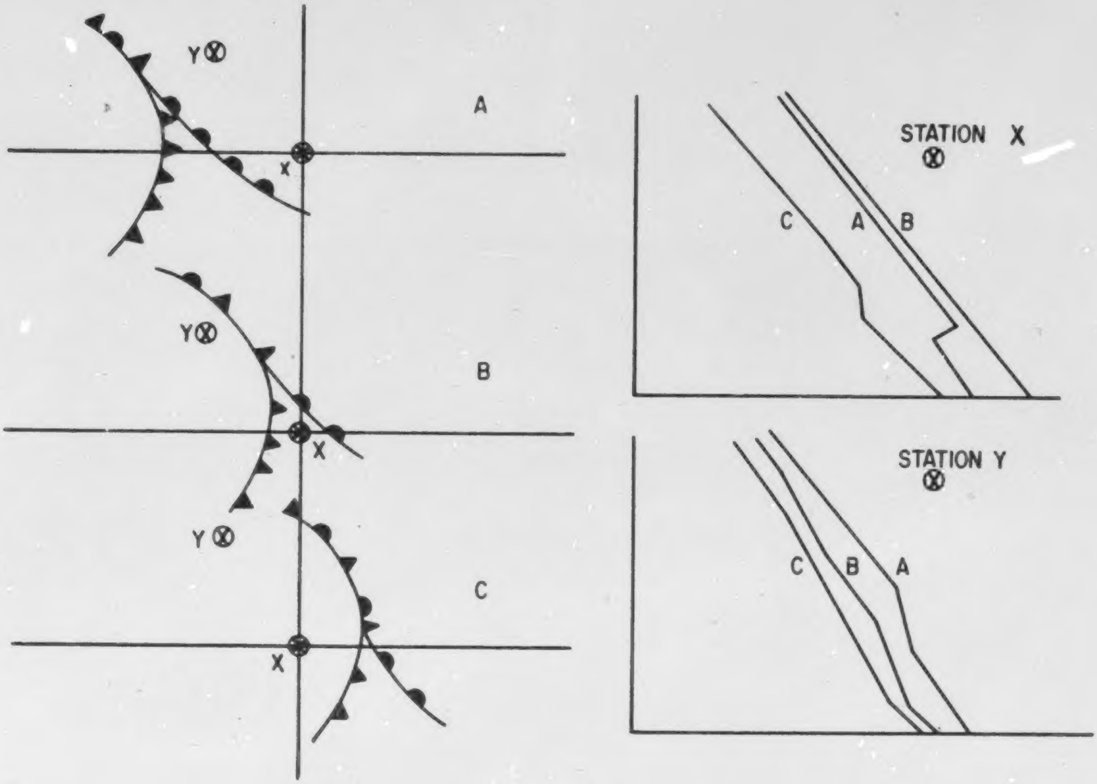

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Figure 5-39.—Illustration of typical lapse rate changes in a cold front occlusion.

Cold front type occlusions are quite common in the western parts of oceans in winter. In summer this type is the rule along the west coast of the United States, Canada, and Alaska where the ocean air is colder than the warm air over the continents.

WARM TYPE OCCLUSIONS

Figure 5-40(A) and (B) illustrates the vertical and horizontal structure of a warm type occlusion in its early stages of development. Line A-A^1 corresponds in each figure.

Figure 5-40(A) shows the structure of the air masses, and figure 5-40(B) shows how the occlusion would be depicted on a surface chart. The occlusion is represented as a continuation of the warm front. The cold front aloft is represented on all charts below the level at which it intersects the warm front.

Weather

The weather associated with warm front occlusions has the characteristics of both warm and cold fronts. The sequence of clouds ahead of the occlusion is similar to the sequence of clouds ahead of a warm front, while the cold front weather occurs near the upper cold front. If either the warm or cool air which is lifted is moist and unstable, showers and sometimes thunderstorms may develop. Weather conditions change rapidly in occlusions, and are usually most severe during the initial stages. However, when the warm air is lifted to higher and higher altitudes, the weather activity diminishes.

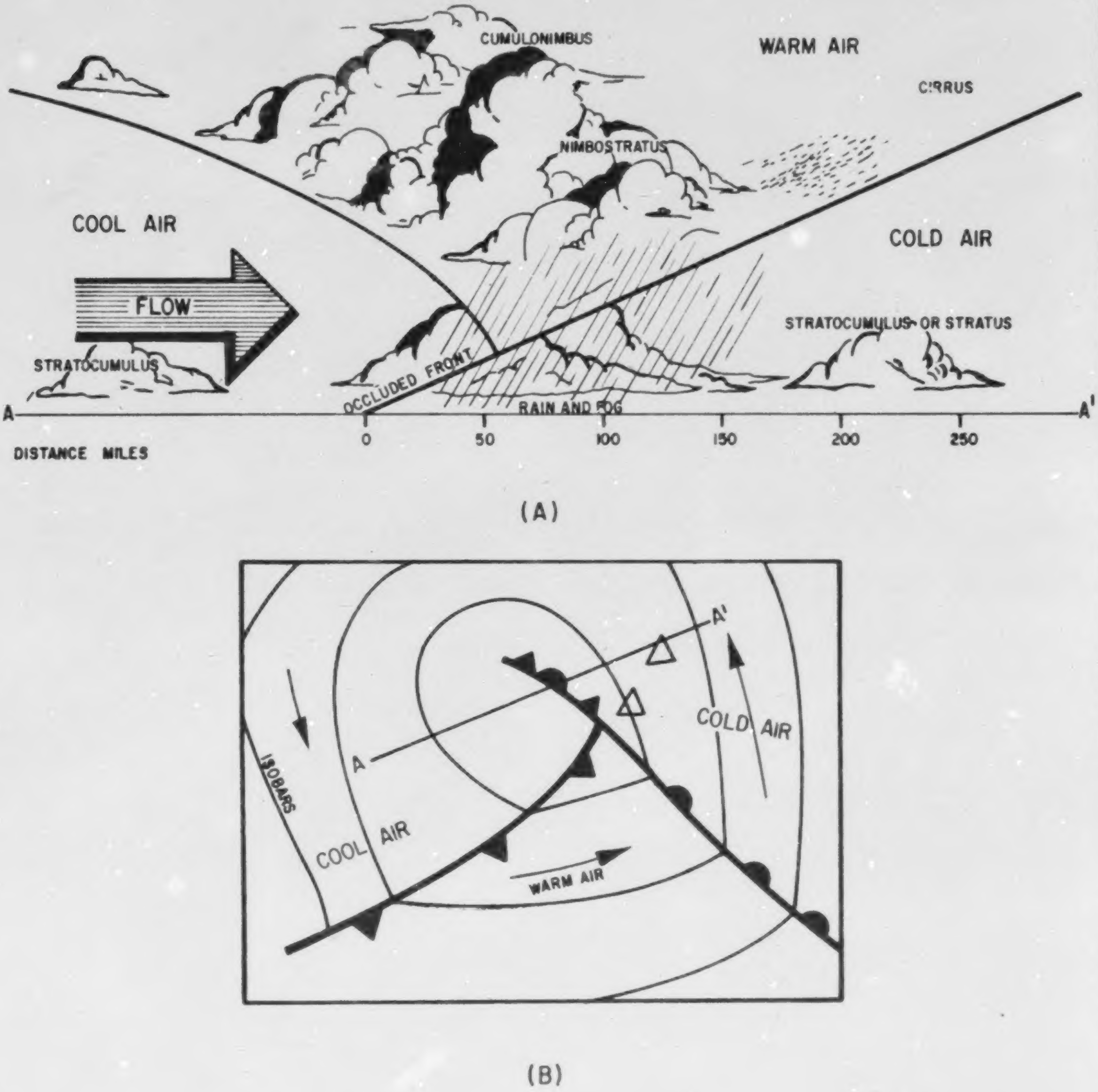
CUMULONIMBUS
WARM AIR
CIRRUS
NIMBOSTRATUS
COOL AIR
COLD AIR
FLOW
STRATOCUMULUS OR STRATUS
STRATOCUMULUS
OCCLUDED FRONT
RAIN AND FOG
A
A'
DISTANCE MILES
0
50
100
150
200
250
(A)
A'
A
COLD AIR
ISOBARS
COOL AIR
WARM AIR
(B)
AG.494

Figure 5-40.—Illustration of warm type occlusion. (A) Vertical structure; (B) horizontal structure.

Showers and thunderstorms, when they occur, are found just ahead and with the upper cold front. Normally, there is clearing weather after passage of the upper front, but this is not always the case.

Surface Indications

The warm type occlusion lies in the same type of pressure pattern as the cold type occlusion.

The most reliable identifying characteristics of the upper front are:

1. A line of marked cold frontal precipitation and cloud types superimposed on warm frontal types ahead of the occluded front.
2. A slight but distinct pressure trough.
3. A line of pressure tendency discontinuities. The pressure tendency shows a steady fall ahead of the upper cold front aloft and, with passage, a leveling off for a short period of time. Another slight fall is evident with the approach of the surface position of the occlusion. After passage the pressure shows a steady rise.

The pressure trough and the accompanying isobar kinks are often more distinct at the upper front than they are at the surface occluded front. Typical isobaric patterns, tendency fields, winds, and weather distribution are shown in figure 5-41.

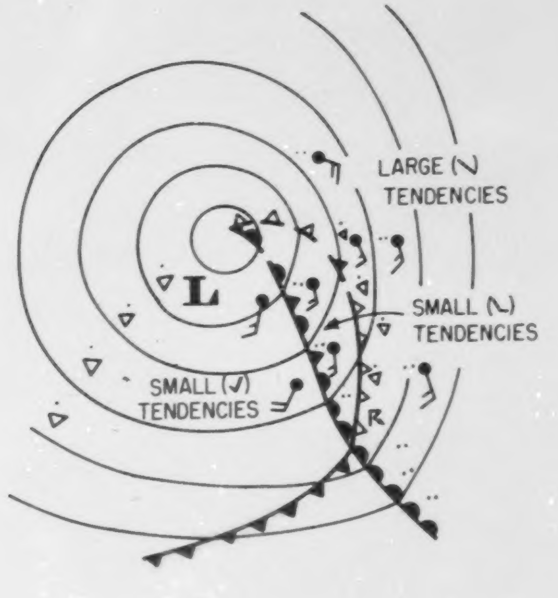

Figure 5-41.—Typical warm front type occlusion on a surface map.

The intensity of the weather along the upper front decreases with distance from the peak of the warm sector, and both fronts become progressively weaker and most diffused as they approach the center of low pressure. Note that the closed isobars are very nearly circular; this is typical of all mature occlusions undistorted by orographic barriers. It is often noted that after occlusion the center of lowest pressure tends to roll back along the occlusion.

Upper Air Characteristics

TEMPERATURE CURVES.—The same statements as those made for cold type occlusions apply to the warm type as well.

UPPER AIR CHARTS.—The warm type occlusion (like the cold type) would appear on upper air charts at approximately the same levels. However, one distinct difference does appear in the location of the warm ridge of air associated with occlusions. The warm tongue, as evidenced by the 1,000-500 mb thickness analysis, reveals that the warm ridge in the thickness pattern lies just to the rear of the occlusion at the peak of its development.

In the late stages of development of both types of occlusions, the warm tongue becomes quite narrow as it works its way around the surface low center. When this occurs, there is usually a closed thickness minimum south of the sea level low.

MODIFICATIONS OF FRONTS

Up to this time we have been discussing typical characteristics of fronts with little regard to the modifications they undergo when passing over certain types of land areas such as mountain barriers. Mountain barriers are but one of the factors that have considerable effect on frontal characteristics. These barriers affect the speed, slope, and weather activity associated with the front. The degree of modification that a mountain barrier has on a front depends on the size of the barrier, the orientation of the front, and the stability of the air masses involved.

MODIFICATION OF WARM FRONTS

On the windward side of the mountain, the precipitation area is widened and the duration and intensity are increased. This is due to slowing down of the front, caused by trapping the cold air between the frontal surface and the mountain. It is also due to increased lifting afforded by the mountains. On the lee side of the mountain, the precipitation area and intensity are decreased.

The front may move over the lee side of the mountain for a considerable distance before reaching the ground again if the air on the lee side is extremely cold.

MODIFICATION OF COLD FRONTS

On the windward side of the mountain barrier, the precipitation ahead of the front is increased due to lifting of the air over the mountains. The speed of movement of the front is decreased, since the air must push up over the retarding barrier. The frontal slope may be steepened as it passes over the barrier; and if the cold front has a higher potential temperature than the valley over which it passes, it may not touch the ground. In this case it will appear as an upper cold front, and can be detected by surface tendencies and associated weather. This situation is common over the Rockies in winter.

On the lee side of the mountains, the movement of the front is accelerated. The intensity of the front decreases due to adiabatic warming as it descends the lee of the barrier. The frontal activity again increases as the front moves away from the barrier.

Wave formation on a cold front may occur when it passes over a mountain barrier if a portion of the cold front is retarded while another portion is not.

MODIFICATION OF OCCLUSIONS

Occluded fronts are affected in much the same manner as cold and warm fronts. Cold front type occlusions usually act as cold fronts, and warm front type occlusions as warm fronts. Mountain barriers, however, may accelerate the occlusion process in that they retard the warm front so the cold front overtakes it more rapidly than if the warm front continued to move at its normal rate of speed.

CHAPTER 6

SURFACE WEATHER MAP ANALYSIS

The Aerographer's Mate, First Class or Chief, who is given the job of forecasting the weather is faced with innumerable problems. As the number of reporting stations and the area covered by these stations increase, the success of the meteorologist depends to a greater extent upon his ability to evaluate all the data. Because more information is available at the surface, the sea level chart continues to be the basis upon which other analyses are based. The higher level analyses, as pointed out previously, aid in constructing a 3-dimensional model of the surface analysis as well as provide charts for other meteorological purposes.

OBJECTIVES OF MAP ANALYSIS

The problem of the analyst is a multiple one. The analyst is faced with a chart upon which is plotted a maze of synchronously observed data whereby he is required to perform the following tasks.

1. Delineate by accepted standard sets of lines and/or symbols the current state of the atmosphere and how it arrived at that state.

2. Decide what particular atmospheric processes are involved in the production of the various kinds of weather reported, representing the processes where possible by accepted standard 3-dimensional models.

3. Complete the analysis in the shortest amount of time so that it may be used for briefing and forecasting purposes.

SUBDIVISION OF THE PROBLEM

As a means of approaching the subject of map analysis the procedure should be divided into logical steps. In this chapter the following approach is used. These categories are listed below in increasing order of extent and importance.

Local Analysis

Local analysis is concerned with the examination of data in a small area or at a fixed spot over a short interval of time such as:

1. Detection of errors in a particular report by comparison with surrounding reports.
2. Location of a front between two adjoining reports.
3. Elimination of unrepresentative elements of a report.
4. Examination of hourly sequences from airports or 3-hourly sequences from weather ships for indications of frontal passage or of inconsistencies between successive reports.

Intermediate Analysis

The scope of intermediate analysis is comprised principally of the fitting of synoptic models to data covering an area; roughly that of a cyclone or anticyclone. The models of troughs, ridges, fronts, wave cyclones, cutoff lows, blocking highs, and occluded cyclones are elements of intermediate analysis. However, such details as the kink in an isobar at a front in an occluded cyclone would be an item of local analysis. Where considerations of intermediate analysis cannot be reconciled with those of local analysis, intermediate analysis should take precedence.

Extended Analysis

The most comprehensive and important category is that of extended analysis. It includes all time and space relationships among the elements of intermediate analysis.

Time relationships are generally grouped under the heading of history or historical sequence. The term continuity is also used with this connotation. If any one element of extended analysis is to be singled out for precedence over all others, it must be history or continuity. This means that no particular weather chart can properly be analyzed in isolation from its predecessors. The first and most important step in the analysis of any weather chart is therefore the transfer of the previous positions and intensities of all significant meteorological fronts and pressure systems of the last analyzed chart to the current unanalyzed chart. This is best accomplished by tracing a line (over a light table) with a yellow pencil or a narrow dashed line with ink, the previous positions of fronts, troughs, and ridge lines, and by spotting the previous positions and central pressures of closed lows and highs. Placing past positions on current charts essentially converts time relationships into space relationships, and subsequent analyses can be carried out in terms of consistent displacements. This particular feature of analysis is extremely important.

Of almost equal importance are the relationships among the various synoptic models on a particular chart and the geometrical and meteorological consistency of analyses at one level with those at neighboring levels in space. Only then can a proper 3-dimensional picture of the current structure of the atmosphere be obtained.

It should be emphasized that the discussion of relative precedence in this section implies correct analyses in each category. Obviously, incorrect extended analysis should never take precedence over correct intermediate or local analysis. For instance, it is often necessary to reanalyze a chart in the light of later data and events shown on later charts.

SUMMARY

You can readily see that a knowledge of the basic meteorological concepts and models presented in earlier chapters of this training manual is a natural prerequisite to a thorough, complete, and correct analysis of a weather chart. Given a workable vocabulary of synoptic models, you should encounter the least difficulty with intermediate analysis. Local analysis will require an understanding and familiarity with the quality and peculiarities of reports. You will become proficient in extended analysis only after you have acquired a basic understanding of the physical and dynamic processes which give rise to the circulation patterns of the atmosphere combined with sufficient experience and knowledge of synoptic charts.

ANALYSIS APPROACH

The surface synoptic weather chart is the principal working tool of the forecaster. The first step in the analysis of data after it has been plotted on the surface weather chart is for the forecaster to be completely familiar with the quality of the data. Knowledge of the probable errors and the types of nonrepresentative observations give the forecaster a better chance of filtering out errors by the smoothing process. The types of errors that can frequently be corrected prior to analysis are: continuing pressure errors from successive reports of ships; communications garbles or errors; pressure tendencies which do not agree with reported pressure; and ship winds which are not in agreement with the magnitude or direction of sea swell or have been plotted in the wrong quadrant of the globe.

When all available synoptic reports have been correctly and completely entered and their quality evaluated, the weather chart is ready for analysis. A number of different recommended procedures are available for the analyst to use. In this chapter, in general, we use the procedure outlined in the first portion of this chapter.

EVALUATION OF DATA

There is always the possibility of an error in plotted reports because of faulty instrumentation, observation, transmission, or charting. Also, even though correct, surface reports are sometimes not representative of the surrounding area or of conditions aloft. The analyst is therefore required to be constantly on the alert for errors or nonrepresentativeness; the quality and usefulness of the analysis depending upon his judgement and keeness of perception when interpreting reports. Location errors are rare for land stations where the station number is printed on the chart. Reports from ships, however, are frequently found to be in error for one or more of the following reasons. Erroneous pressures are due to inaccurate reading or calibration of the barometer and by either applying corrections incorrectly or by failing to apply the needed corrections at all. Errors in transmission by radio, errors caused by difference in time, and frequent errors of ship's position by 5° to 10°, either in latitude or longitude. A good remedy for the position and other type errors from ships is to keep a running log of reports from merchant ships. On the log, successive reports from any given ship may be plotted side by side for ready comparison.

Diurnal Variation of Meteorological Elements

Processing of large numbers of observations for each hour of the day has shown that a certain kind and amount of change in any meteorological variable can be expected simply because of the varying times of day of the observation. This change is called the diurnal variation of the elements. There are also considerable seasonal differences in the elements. For example, along coastal stations, the onset of the sea breeze in the late morning hours and a subsequent maximum velocity in the early or middle afternoon is especially evident at midlatitude and some tropical stations. Over land, the time of maximum wind velocity is usually early afternoon when thermal convection has reached its maximum and the time of minimum is the early morning hours when the atmosphere is most stable. Similar variations are apparent in pressure tendency, temperature, clouds, and precipitation.

Temperature varies much less over oceans than over land diurnally due to the small daily change of the ocean surface temperature. In general, as the distance from water bodies increases, the average diurnal temperature range also increases. Precipitation at most inland stations and more particularly in the Tropics shows a maximum in the afternoon coupled with the maximum of convectional storms at that time. At tropical islands, oceans, and coastal stations, the maximum precipitation appears at night.

Detection of Erroneous Data

The validity of a report or part of a report becomes suspect when it is inconsistent with nearby reports (in case of dense synoptic networks), contains internal inconsistencies, or leads to marked or unlikely changes in continuity or history (in areas of sparse data).

The first two cases involve principally a problem of local analysis. The reports should be compared with neighboring data, exercising care that comparative data are observed at the same altitude and over the same type of underlying surface. The last case involves an isolated report and requires the application of all categories of analytical tools. If possible the previous 3-hourly reports from this station should be consulted to check for continuity. Sometimes in the case of an apparently erroneous isolated report it is virtually impossible to check its validity. However, never disregard or discard an isolated report simply because it is difficult to fit into a preconceived pattern. At least one or more valid logical reasons should exist for not drawing to such a report. Many analysts have violated this rule only to find that subsequent charts confirmed the existence of an important meteorological event.

Climatology can also be used to detect erroneous reports; for example, snow in the summer or an 80°F temperature inside the Arctic circle. Random errors are the most difficult to detect unless they are large, while systematic or repeated errors are more easily discovered and corrected.

Classification and Correction of Errors

Once an error is detected, the data must not be discarded arbitrarily. Some attempt should be made to estimate the proper correction of the data. After detecting the error, you should make a careful consideration of the possible sources of the error. A convenient classification of errors by source is the following: errors due to encoding, transmission, decoding, and plotting; errors of observations; and computational errors in data not directly observed.

Errors in the first group are essentially communications errors. Mistakes in decoding or plotting are the most easily checked, requiring only a re-examination of the original message. This should be the first step toward correction of inconsistent or suspect data. A transposition of numerical digits for the correct ones can occur both in landline operation and more commonly in CW messages which are in Morse code. Substitutions of a 1 for a 6, or 2 for 7 can occur in the CW message if the operator transposed dots and dashes. Frequently, on teletypes the neighboring digit is substituted for the correct one, for example a 5 or a 7 for a 6. These errors are equally probable in any numerical group, but only those occurring at positions where the units being transmitted are fairly large will they be detectable and correctable. Thus, errors in the tens digit of pressure or temperature are usually obvious, but similar transpositions in the units digit cannot be identified so easily. Errors of this type are also common in wind direction and ship position reports.

As the number of errors or inconsistencies in a single report increases, so does the probability that the remainder of the report is also erroneous. In the analysis of these, as well as other communications errors, a well-trained, experienced, and conscientious plotter is of inestimable value to the analyst. However, even an experienced plotter should never be allowed to change any element of a report. This is a job for the analyst. The plotter should be given instructions to distinguish these types of data, such as placing a question mark next to the dubious element or report.

Observational and computational errors of most reports received from other weather units are almost impossible to correct. However, some stations may make consistent, obvious errors in weather element values and a list of these stations should be maintained with the probable correction to be applied to the weather report.

Computational errors are more likely in upper air data than sea level data, since most of the latter are directly observed. Exceptions are sea level pressure reductions, dewpoint temperatures, and true wind speed and direction reported by moving ships. The first two elements are generally taken directly from tables or graphs prepared especially for such calculations and the errors arise principally from the use of incorrect or unrepresentative entries to the tables or graphs. Greater reliance should be placed on true wind reports from ships known to carry well-trained observers, such as weather ships, naval vessels, and large commercial liners. Many smaller vessels are not equipped with anemometers, in which case direct estimates of true wind are made using the Beaufort scale to correlate the state of sea with the wind speed.

REPRESENTATIVENESS OF DATA

Even when all the probable errors in a report are accounted for, parts of it may still be inconsistent with corresponding parts of other reports nearby or within the same air mass. Such a report is then said to be unrepresentative of the property being delineated. Any meteorological element subject to purely local influences such as heating or cooling, terrain, water sources, local convergence, and the like, is likely to be unrepresentative. Except in the vicinity of well-defined currents, ocean surfaces are more homogeneous horizontally than are land surfaces, so one would expect reports from ships and small vessels to be more representative than continental reports. This is generally true, but important exceptions are noted. Similarly, properties of air in the friction layer are more subject to local influences than are the corresponding elements of the free atmosphere. Here is the principal reason for the continual insistence that no surface analysis can be considered completed until it has been shown to be geometrically and meteorologically consistent with corresponding analyses in the free atmosphere. Since surface or sea level data are observed at the base of the friction layer, where local influences are most effective, it is easily possible to find conditions under which each element of the surface report is nonrepresentative. The following section contains a discussion of the representativeness of some of the various meteorological elements.

Sea Level Pressure (PPP)

Pressure at station level is, by definition representative of the mass of the air column or unit cross section above the station. Station pressures at different levels must be reduced to a common reference level if they are to be useful. The most common level for surface charts is sea level. Sea level pressure reported by a station not at sea level is a convenient fiction arrived at by substituting for the missing part of the air column a column of nonexistent air with an estimated mean density and of length equal to the altitude of the station. The representativeness of such derived pressure is thus determined by the representativeness of the mass of the substitute air column. Obviously, the shorter the column, the greater is the chance of its having a representative mass. Overestimates of the mean temperature of the column in question will result in reported sea level pressures that are unrepresentative on the low side and vice versa. Thus, a mountain station with a temperature lower than its neighboring stations will report a higher sea level pressure. This is also the reason why the intensity of thermal highs and lows in mountain and plateau areas is, to some extent, fictitious on sea level charts. It is also the reason why some other level than sea level, as the reference level, is used for constructing charts in these areas.

Pressures reported by stations at or near the reference level are the most representative of all the meteorological elements, although (especially in the case of ship reports) they are subject to error.

Sea level pressure reports from mountain stations which extend well above the average surrounding terrain should be disregarded in drawing the sea level pressure pattern. For example, a station such as Leadville, Colorado, is 10,158 feet above sea level. Since the pressure at 10,000 feet is about 700 milibars, the correction to sea level is about 300 milibars. You can readily see that an error of only 1 percent in estimating the mean density of the substitute column would therefore result in a 3-mb error in sea level pressure. Station level pressure, or sea level pressure at a station near sea level, would be the most representative element of the meteorological report.

Table 6-1 shows probable errors in pressure observations that are used by the National Meteorological Center (NMC).

Pressure Tendency and Net 3 Hour Change (app)

Ships without a barograph do not report pressure tendency. Land stations report the most accurate 3-hourly changes in pressure, followed by the stationary ships such as Ocean Station Vessels. Some shipping is subject to erratic course and speed variations which make pressure tendencies and net 3-hour changes seem inaccurate.

The characteristic of the change in pressure and the amount of the change are very important clues to developing weather situations. On a moving ship, the pressure change indicated by the barograph is due to the actual change in atmospheric pressure plus the change in pressure as the ship moves in relation to the high- and low-pressure areas. For example, a ship sailing eastward at 15 kt and being overtaken by a low-pressure system moving at 20 kt would show a slowly falling pressure characteristic. Thus, two ships in the same area might actually report different pressure characteristics and changes in the $D_sv_s$app group. The forecaster is able to correct apparent pressure change to true pressure change by using the direction of movement and speed of the ship (D_sv_s), and the movement of the pressure system.

Temperature (TT)

Any process or condition which tends to produce cooling in the lowest layers of the atmosphere will cause a low-level or ground inversion, the intensity of which is a measure of the resultant unrepresentativeness of the surface temperature. The most common of such processes and conditions are (1) nocturnal radiation, (2) advection over a colder surface, (3) drainage

Parameter	Probable Error	Remarks
Reduced sea level pressure (Land stations with calibrated mercurial barometers.)	± 0.5 mb per 1,000 ft elevation	This error results from nonrepresentative temperature parameter in the reduction to sea level computations.
Weather ship pressures	1 mb	Occasionally a weather ship will leave port with its barometer miscalibrated.
Commercial ship pressures	± 1 to 2 mb	This error is frequently constant with a given ship. Most ships use aneroid barometers, which are less accurate than mercurial barometers.

Table 6-1.—Errors in pressure observations.

of cold air into valleys, (4) snow on the ground, and (5) evaporation from a local water source. All of these processes except (2) and (5) are ineffective or irrelevant at sea, so temperatures are generally more representative at sea than on land. The prevalence of these processes on land makes temperature the least representative of all elements of the surface report from land stations.

Table 6-2 summarizes the relative unrepresentativeness of temperature under various conditions. On land, surface temperature is in accordance with table 6-2.

In winter a snow cover chart should be used as an aid in evaluating surface temperatures. The edge of the snow field often appears in the temperature field as a pseudo front.

Two exceptions to the representativeness of surface temperature at sea should be noted: (1) Marked ocean currents affect air temperatures over or near them; and (2) the internal warmth of most modern ships renders the proper exposure of thermometers almost impossible. Even with the best exposure, ship temperatures appear to be about 1°F too high.

Because the amount of unrepresentativeness of surface temperature in a warm stable air mass may exceed the frontal contrast, the curious case in which the temperature rises following a cold frontal passage can occur. Typical vertical temperature distributions in such a situation are shown in figure 6-1. The solid line with ground inversion represents conditions before the front passed, and the broken line with the higher frontal inversion represents the post frontal condition.

Frontal temperature contrasts can also be masked on night and early morning charts if the warm air mass is clear or nearly cloudless, with light winds, permitting a marked radiation inversion to form. Free atmosphere temperatures should then be used in locating the front discontinuity, for example, on the 850-mb chart, where this surface is not too near ground.

Stations at high altitudes with unrepresentative temperatures result in incorrect reductions of pressure to sea level. High level stations under clear conditions at night and early morning report sea level pressures too high, resulting in a fictitious high in the surface analysis. Such a

Most representative	Unrepresentative
On afternoon maps.	On early morning maps.
*In unstable (cold) air masses.	*In stable (warm) air masses.
*In areas of strong winds.	*In areas of calm or light winds.
Windward side of mountains.	Leeward side of mountains.
On equatorward slopes of hills and mountains.	At valley bottoms and high peaks.
*When no precipitation is occurring.	*During precipitation.–
In cloudy areas at night.	In clear areas at night.
On windward sides of lakes.	On leeward sides of lakes.
Over grass-covered areas.	In deserts, forests, and snow fields.
Statements marked with (*) are equally true of ship reports.	

Table 6-2.–Surface temperature on land.

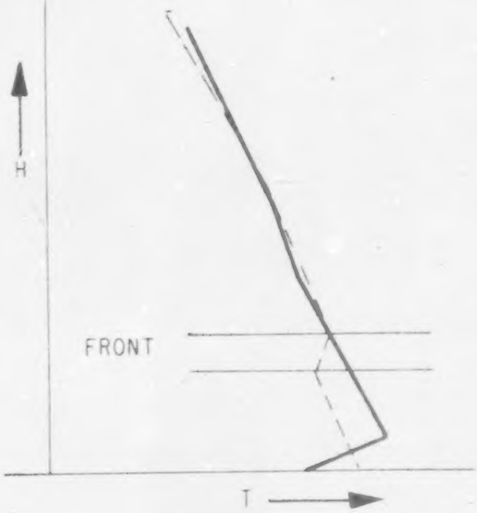

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Figure 6-1.–Illustration of a temperature rise with a cold front.

fictitious high should be avoided by reference to the wind field just above the surface friction layers. The centers of clockwise circulation at these levels (in the Northern Hemisphere) accurately pinpoint the existence of true high pressure centers. Where pressure reductions are representative, the high center and the center of clockwise circulation are generally coincident.

Dewpoint ($T_d T_d$)

The dewpoint temperature is representative whenever the air temperature is, and is often representative when air temperature is not. This is due to the fact that dewpoint is relatively unaffected by dry adiabatic and isobaric nonadiabatic processes except those involving evaporation and condensation. At stations near a water source or wherever precipitation is occurring, dewpoint temperature will be unrepresentative; under most other conditions it is more representative than temperature and can often be used to find fronts where the temperature contrast is masked.

The dewpoint discontinuity is often the only means of locating the boundaries of an advancing wedge or tongue of mT air. The so called "dewpoint front" is generally located near the 60-degree dewpoint isotherm in winter. It occurs most often in the Gulf States and can best be identified after it has moved well inland from the gulf coast, because there is always some dewpoint contrast between a coastal and an inland station, due to the water effects at coastal points. The western boundary of this marked dewpoint contrast occurs in Texas and southern

Oklahoma during much of the year, but particularly in warmer seasons. This north-south line oscillates from west to east, but rarely moves east of Dallas. It separates the low dewpoints of dry cT air over the southwest United States from the moist mT air over the southeast United States. Even though it is so persistent as to be almost a climatological feature of this region, it is not a true front. Since fog, drizzle, low stratus, and restricted visibilities are most likely to occur in this moist tongue, it is a forecasting necessity to keep track of its boundaries. Shading the whole tongue light red is an analytical convenience which avoids the error of indicating it as a front.

Elevation differences between neighboring stations in the same air mass generally show a much smaller difference in dewpoint than in air temperature. This is due to the fact that the dewpoint temperature diminishes vertically at only about one-fifth the rate of decrease of air temperature.

For the reasons outlined above, the dewpoint temperature is one of the most reliable and useful elements of the surface report.

Wind Direction and Speed (ddff)

Ocean wind reports from vessels are in general reliable, but those from land stations are often affected by local topography, making them unrepresentative for a short distance above the surface. This zone is called the friction layer. Wind in the friction layer is much more subject to local influences over land than over water. The principal causes are terrain, vegetative cover, and local heating and cooling. In many places terrain permits air to move only in certain directions with respect to its own orientation. It also acts as a windbreak for points on its lee side. Frictional drag varies with vegetative cover, and local heating deforms the pressure pattern by realining the surface isobars more nearly parallel to the surface isotherms. For all of these reasons, a chart of winds above or near the top of the friction layer is indispensable to accurate surface analysis over land areas. Except in very mountainous areas, the winds 2,000 feet above the surface (not sea level) have been found adequate.

The thermal deformation of local isobaric patterns is most marked along coastlines and lakeshores, and results in adding to the representative wind the so-called sea or land breeze component. Land breeze components are generally smaller and more normal to the shoreline. An estimate of the representative wind can be obtained by subtracting the sea or land breeze component vectorially from the (nonrepresentative) observed wind, or by consulting wind reports at higher levels. The winds reported from small islands often show evidence of sea breeze components.

Local convective activity can affect surface winds to a marked degree in the immediate vicinity because of the extreme local convergence and divergence necessary to maintain or compensate for the vertical motion.

Table 6-3 shows the reliability of wind direction and speed over land surfaces.

All the factors discussed in table 6-3 affect both the speed and direction of the wind. There is further important correlation between these two properties of air motion: direction is most representative when speed is 10 knots or greater. Unless there is reason to believe that it also is unrepresentative, pressure should be given precedence over wind direction in drawing isobars in areas of light winds.

A physical tool which is commonly used over oceans where winds from ships can frequently be converted into a gradient wind level is the geostrophic wind scale. This scale is discussed later is this chapter. Studies at the NMC, and by others, show that by increasing the surface wind speed by one-third and adding 20° to its direction, a good estimate of the sea level pressure gradient results from ship reports. Although this method is not completely accurate, this procedure in estimating the horizontal pressure gradient is a valuable tool in constructing analyses over sparse oceanic areas.

Present Weather (ww)

The weather occurring at a station is most likely to be representative if it extends over an appreciable length of time, but even intermittent precipitation can be scattered thoughout an air mass and hence would be representative. Once the possibility of error has been eliminated, it is

Most representative	Unrepresentative
In open flat areas.	At coastal or island stations.
	At mountain or valley stations.
Over bare soil.	In forested areas.
Windward side of orographic obstructions.	Leeward side of orographic obstructions.
*In unstable air masses.	*In stable air masses.
In daytime.	At night.
	*Near ⧗ , ▽ , ☈ , etc.

Items marked (*) apply equally to winds at sea.

Table 6-3.–Reliability of wind data.

almost always necessary to assume that reported present weather is representative.

Smoke, dust, sand, haze, and fog are often purely local phenomena, and hence can be unrepresentative. However, fog and dust can be characteristic of a large part of an air mass, particularly advection fog and the dust storms of southwestern United States.

The time of observation, with respect to the diurnal maximum of various types of weather, should also be considered.

Clouds (C_L, C_M, C_H)

High and middle clouds are more likely to be representative than low clouds or clouds with great vertical development. As in the case of present weather, the diurnal variation of the latter types is important in the evaluation of their representativeness.

A striking example of unrepresentative low clouds is the west-coast stratus, which rarely extends inland for any appreciable distance.

Visibility (vv)

Only restricted visibilities are likely to be unrepresentative. Such visibilities are associated with unrepresentative present weather discussed in a previous section. Low visibilities are not likely to be the result of errors of judgment, due to the existence of markers at accurately known distances and instrumental observations at a land station. Over the sea, visibility may be quite unrepresentative due to the absence of such markers or instrumental means.

GENERAL SURFACE ANALYSIS PROCEDURE

There are several factors which influence the order in which a map analyst draws the various elements of his analysis on the surface weather map. These factors include delayed receipt of reports, type of map and impending weather, historical sequences available, relative difficulty of analyses in specific areas, auxiliary upper level data available, time required for completion, and skill and experience of the analyst. Moreover, very often the various elements of the analysis are drawn concurrently or certain elements are sketched in lightly (or mentally) before other elements are drawn in their final form.

In general, the various steps and elements of the analysis are usually carried out in a manner similar to the following outline:

1. Indicate in ink (or yellow pencil) at least three previous positions of all centers which are expected to be found on the current chart. Also, when available, draw in dashed ink (or yellow pencil) at least two previous positions of all fronts which are expected to be carried forward to the current chart.

2. Try to delineate fronts before drawing isobars. This is not always possible, especially when the front is weak or diffuse. In this case, sketch in isobars in the areas where the front is most probable.

3. Draw isobars or continue isobaric analysis to delineate highs, lows, and other features of the pressure pattern.

4. Illustrate by color the sketched-in positions of the fronts, keeping in mind that the color of the front is determined by the instantaneous motion of the cold air mass.

5. Print a red "L" near the center of each cyclone and a blue "H" at the center of each anticyclone.

6. Label air masses, if appropriate.

7. Color in all present and past weather areas, as appropriate.

8. Draw isallobars—this is optional and depends upon local policy.

The order may be varied to conform to local requirements and other factors as mentioned above.

ISOBARIC ANALYSIS

At every point on an isobar the barometric pressure, reduced to sea level, is the same. If a free air map is drawn on a vertical cross section and points of equal pressure are connected, a line of constant pressure is found which represents the isobaric surface. The surface is usually not a flat surface but will have hills and valleys.

Since an accurate surface analysis is an important supplement to upper air data in determining many details of analysis of upper air charts, the final surface analysis must also reflect general upper air conditions.

Before proceeding with the analysis, the analyst should keep in mind the following considerations:

1. Compatibility of the analysis with both current data and continuity, so that meteorological events portrayed on the current chart show logical development from previous charts.

2. Consistency between the actual windfields and the movement of fronts and air masses.

3. Regard for reasonable continuity on the basis of accepted models or experience in indicating intensification or weakening of pressure systems, where data are inadequate to directly determine their intensity.

4. Care in showing the isobaric angle at a front, especially to avoid violation of data in attempting to conform to a preconceived model.

All three categories of analysis are applicable in the drawing of isobars. Local analysis is used in interpolating between two reports, neither of which has the exact value to the particular isobar being drawn and it is also used in deciding which way the isobars are to be drawn around cols, troughs, ridges, etc. Intermediate analysis provides the synoptic models. Extended analysis in the form of history or continuity is used in obtaining an approximate position and general shape of closed centers. Relating cols to closed centers by drawing intersecting trough and ridge lines is also extended analysis.

ANGLE OF THE WIND WITH THE ISOBARS

Owing to friction, the wind does not blow along the isobars exactly but a little left or across the isobar from higher to lower pressure. Because of friction between the air and sea or land surfaces, the wind will blow across the isobar from higher to lower pressures at an angle of 10° to 20° over sea; the greater the wind speed the greater is the angle of the wind direction to the isobar. There are exceptions to this rule however, especially over land, and you should not insist too rigidly on this rule. Figure 6-2 illustrates the influence of the earth's rotation and friction of the wind direction.

WIND SPEED AND SPACING OF ISOBARS

From the discussion of the basic wind theory in chapter 4, several relationships between wind speed and isobar spacing are evident:

1. The spacing of isobars is inversely proportional to the wind speed. Isobar spacing is a representation of pressure gradient, and pressure gradient is directly related to wind speed. Winds are strong in areas of large/strong pressure gradient and isobars are close together; winds are weak in areas of small/weak pressure gradient and isobars are far apart.

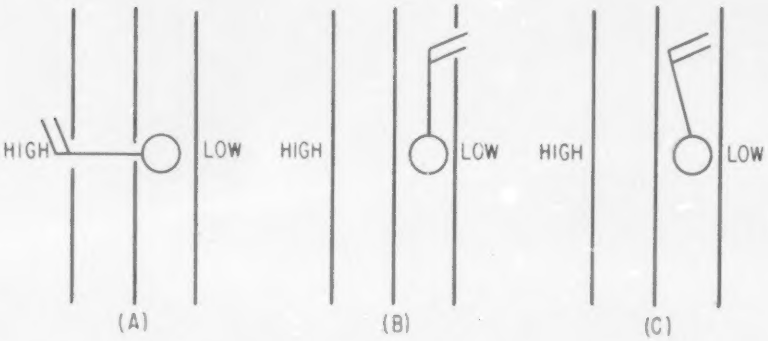

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Figure 6-2.—Angle of wind direction to isobars. (A) If the earth did not rotate; (B) influence of earth's rotation; and (C) influence of both earth's rotation and friction.

2. For a given wind speed, the spacing between isobars decreases with increasing latitude. Table 6-4 shows the spacing of isobars at 4-mb intervals for geostrophic wind speed versus latitude.

From table 6-4, it may be seen that at latitude 40° with wind of 20 knots, the isobars at intervals of 4 millibars should be separated by 179 miles. At 60° with 20 knots, the isobars should be 133 miles apart.

3. For a given wind speed, the space between isobars will be greater with anticyclonic curvature than with cyclonic curvature. With a given wind speed and straight isobars, the space between isobars will be less than with anticyclonic curvature and more than with cyclonic curvature.

Sometimes in areas where data are scarce, you may have to draw isobars with very little information available. Isobars are crowded more closely together in areas where the wind is strong but farther apart where the wind is light. This principle helps when observations are scanty as illustrated in figure 6-3(A) and (B). When there is a large difference of pressure between two ships on the chart, several isobars must be drawn. If there are no intervening entries it is a good idea to determine the number of isobars to be drawn and space a series of dots between the ships to use as a guide. If one of the ships has a considerably stronger wind than the other, the isobars should be more closely spaced near the ship with the strong wind.

Wind speed observed (knots)	Approximate distance (in nautical miles) between isobars drawn for every 4 millibars			
	30°	40°	50°	60°
10	461	358	301	266
15	307	239	200	177
20	230	179	150	133
25	184	143	120	106
30	154	119	100	89
35	132	102	86	76
40	115	90	75	66
50	92	72	60	53
60	77	60	50	44

Table 6-4.—Geostrophic wind distance between isobars over ocean at 4-mb intervals for various wind speeds and latitudes.

GEOSTROPHIC AND GRADIENT WIND SCALES

Because the pressure gradient is closely related to one of the principal components of the

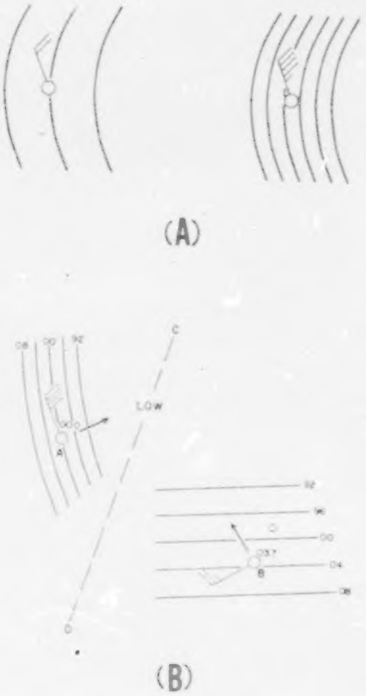

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Figure 6-3.—Isobaric spacing. (A) In accordance with wind speed; (B) in accordance with the geostrophic wind scale.

wind, and in fact determines the general direction of the windflow, a powerful additional tool is afforded the analyst in delineating the pressure field. Where wind observations are available, isobaric analysis is more easily and accurately accomplished than is the analysis of any other scalar quantity of the atmosphere. From Buys Ballot's law, relating wind and pressure, the approximate location of high and low centers can be found without any pressure data whatever, if sufficiently reliable wind data are available.

There are many sizes, types, and shapes of geostrophic wind scales but all have one thing in common—they are either a graphical or tabular solution of the geostrophic wind equation. The computations made on the geostrophic wind scale are for straight isopleths above the friction layer. Gradient wind scales are graphical solutions of the gradient wind equations where the radius of curvature of the path of the air parcel is taken into consideration. Corrections are made for anticyclonic and cyclonic curvature.

One type of scale has as its basic unit of length in construction the degree of latitude. This makes this type scale independent of the scale of a particular map projection and is equally applicable to all charts. The most convenient type scale is one which can be placed on the chart where the wind report is plotted or where wind measurements are to be made. This requires that the wind scale be constructed to fit the map scale, which is constant only on the standard parallels of the projection. Such geostrophic wind scales have been constructed for most meteorological charts, and are either printed in one corner of the chart or on a separate transparent overlay.

In most circumstances the gradient wind is a better approximation to the true wind than is the geostrophic wind. The gradient wind scale should, therefore, be more useful than the geostrophic scale, but in practice it is not, principally because of the complications involved in determining the radius of curvature of the path followed by the air parcel. Since it is such a laborious job to correct for curvature, qualitative estimates of the curvature correction are generally made.

Wind analysis in the free atmosphere is somewhat different. Gradient wind scales which, in effect, compute path curvature from contour curvature and contour movements have been developed for use on upper air charts.

Further details on the geostrophic and gradient wind problem and various scales developed for use by the Navy can be found in Meteorological Wind Scales, NW 50-1P-551.

Use of the Geostrophic Wind Scales

Two types of geostrophic wind scales which may be used for surface analysis are discussed here. The first type is the one that is commonly printed on the base map; the other is an overlay type.

USE OF GEOSTROPHIC WIND SCALE (PRINTED ON BASE MAP).—An illustration of the type of geostrophic wind scale for sea level surface weather maps is shown in figure 6-4.

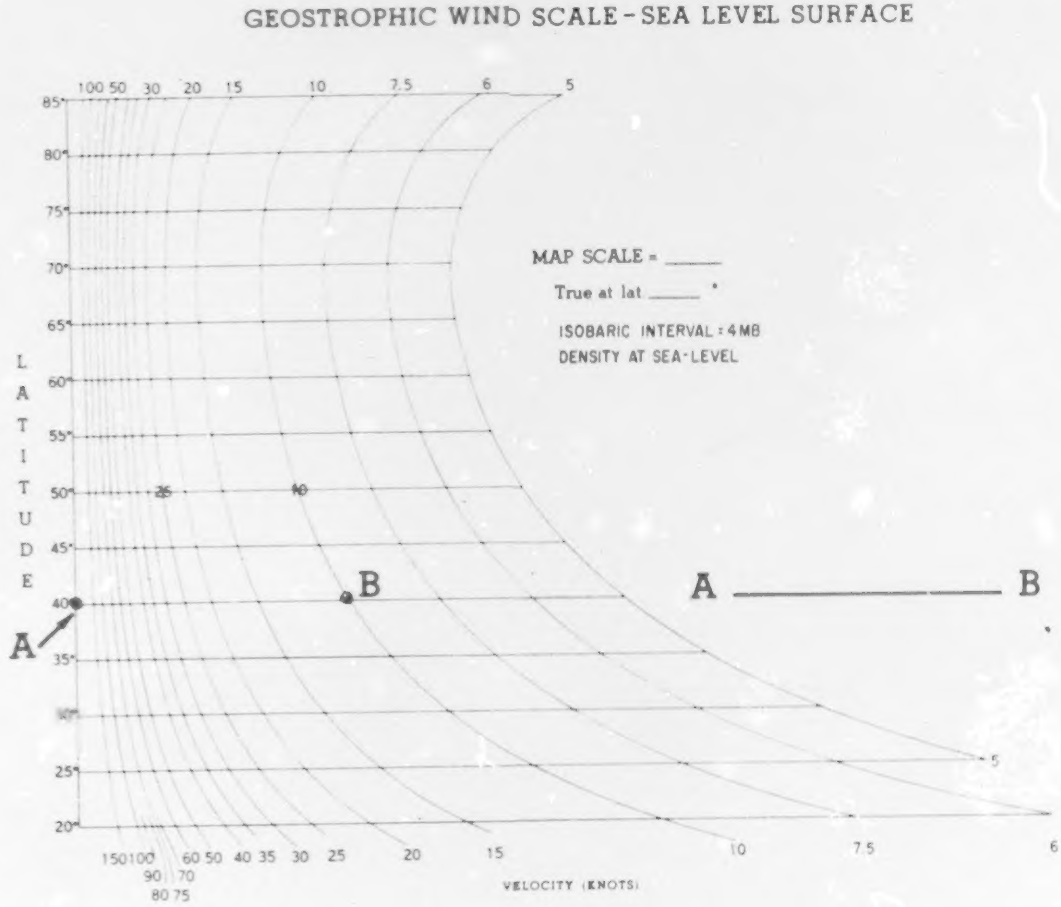

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Figure 6-4.–Use of geostrophic wind scale commonly printed on base maps.

To determine the wind speed at a point over the ocean, you would proceed as follows: First, place the edge of a sheet of paper at right angles to two successive isobars at the place on the weather map where you want to find the wind speed. Mark the distance between the isobars on the edge of the sheet and also note the latitude. Suppose the distance between the isobars as marked on your sheet of paper is equal to the line AB as shown at the right in figure 6-4; also that the line AB was at latitude 40° on your chart. Starting at the left of the scale, measure off the distance AB on the line for latitude 40°. It can be seen that point B coincides with a curved line in the scale. Looking down this curved line from point B, you can then read off at the base of the scale the geostrophic wind speed, which in this case is 10 knots. If point B of the line representing the distance between isobars falls between two curved wind speed lines on the scale, you can obtain the wind speed by interpolation.

If the distance you measured between two isobars on the weather map extended from 35°N to 40°N, for example, you should use the mean of these figures, that is, latitude 37.5° to find the wind speed.

Similarly, you can also obtain from the scale (fig. 6-4) the spacing between two isobars if both the wind speed and latitude are known. As an example, you will use the values for the wind speed (10 knots) and latitude (40°) as in the preceding case. Starting at the bottom of the scale, move up the curved line marked 10 knots until you reach the latitude 40° line, point B. The line BA (fig. 6-4) is the correct spacing for two successive isobars at latitude 40° for a wind speed of 10 knots.

It is also important to remember that the geostrophic wind scale as shown in figure 6-4 is merely an example and should not be used for computing wind speed and the spacing of isobars on your weather charts. You should use the geostrophic wind scale as printed on your map base and draw your isobars for the same intervals as indicated in the scale.

USE OF GEOSTROPHIC WIND SCALE (OVERLAY TYPE). The type wind scale illustrated in figure 6-5 is one of those presented in, Meteorological Wind Scales, NW 50-1P-551. By cutting out the cards indicated in Appendix A of that publication you can obtain wind scales for Polar Stereographic, Lambert Conformal, and Mercator Projection maps. Two cards are available for each standard map scale, one for surface winds and the other for upper air winds.

To use this type scale for surface winds, select the appropriate scale for map type and map scale. For a given latitude, place the latitude number from the card on an isobar line, and count the number of millibars between this latitude line and the INDEX (interpolate where necessary). Multiply the number of millibars by 10 to obtain the geostrophic wind for that isobar spacing at that latitude. (See fig. 6-5.) In the illustration given in figure 6-5 place the 45° mark on the card on an isobar line (1012) and count millibars to the INDEX (2 mb) and multiply by 10 for knots (20 knots).

Observed Versus Geostrophic Wind

The preceding section was devoted chiefly to the spacing of isobars. Their direction should be determined by assuming a certain amount of crossisobar flow toward lower pressure. Observed values of the actual angle between surface wind and isobars in middle latitudes vary from a minimum of 10° over oceans by as much as 45° or more over rough terrain. An average value of 20° over water and 35° over land can be used in the absence of data to the contrary. The frictional effects which cause the cross isobar flow also reduce the speed of the observed wind as compared to the gradient wind. So many factors are involved as to make any really accurate estimate of frictional effects on wind speed virtually impossible. Following the practice used by the NMC, stated previously, will be helpful in determining gradient wind speed and direction over oceans.

FLAT MAPS

On the flat map, winds are light and variable and pressures in certain regions, where isobaric analysis is attempted, are relatively uniform. The wind movement is irregular with no systematic differences between barometric pressure readings to the right or left of the wind. When the map, or any large section of it, is flat it is helpful to study the previous map prepared 12 or 24 hours earlier. A well-defined cyclone on the previous map is likely to persist, in a modified form, and may be possible to identify within the flat section of the current map. Also, keep in mind that there are certain permanent or semipermanent features of the pressure distribution and wind circulation of the oceans such as the Aleutian Low and the Azores High. Even when the map is flat it is usually possible to trace wind circulation and the characteristic, though slight, variations that exist in the vicinity of the centers.

COMPUTER PRODUCTS

Since the advent of computer produced products the accuracy of meteorological analysis and prognosis has increased considerably. The availability of computer products has provided a unique flexibility in the utilization of personnel and raw data. The reliance of the forecaster on computer produced charts is normally proportional to the time frame he is working under and the availability of personnel to plot and analyze by conventional means. In many instances computer products are used entirely, either due to personnel limitations or because the computer product is actually more accurate. The computer is capable of digesting a vast amount of data in a limited amount of time and presenting the

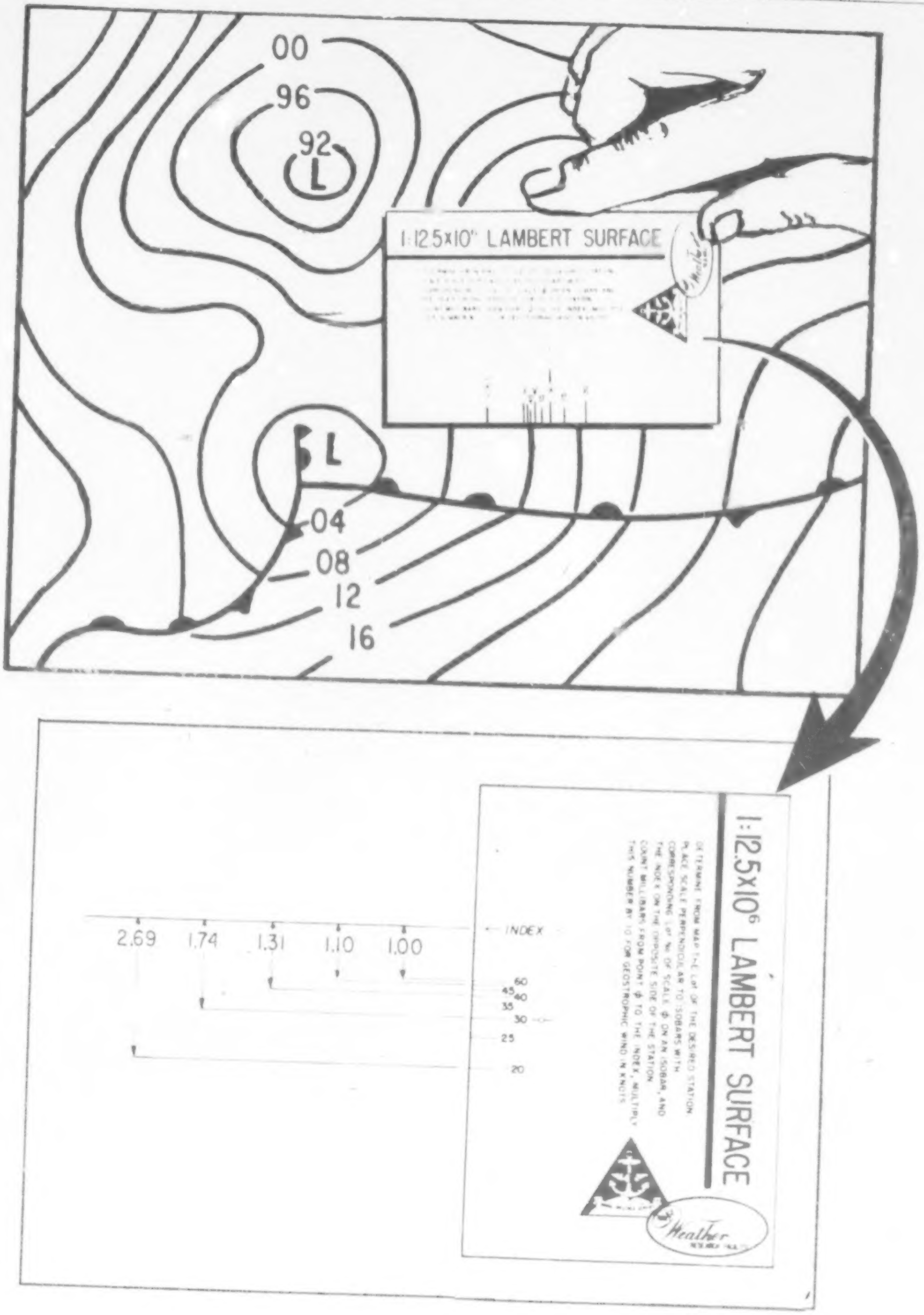
00
96
92
L
L
04
08
12
16
1:12.5x10⁶ LAMBERT SURFACE
DETERMINE FROM MAP THE LAT OF THE DESIRED STATION.
PLACE SCALE PERPENDICULAR TO ISOBARS WITH
CORRESPONDING LAT № OF SCALE ⊕ ON AN ISOBAR, AND
THE INDEX ON THE OPPOSITE SIDE OF THE STATION.
COUNT MILLIBARS FROM POINT ⊕ TO THE INDEX, MULTIPLY
THIS NUMBER BY 10 FOR GEOSTROPHIC WIND IN KNOTS
INDEX
2.69
1.74
1.31
1.10
1.00
60
45
40
35
30
25
20
Weather

Figure 6-5.—Use of geostrophic wind scale (overlay type).

AG.500

forecaster with a finished analysis or prognosis otherwise impossible to achieve. There are many advantages to the machine produced product such as the elimination of many human errors. However, the close attention to detail necessary in performing hand analysis may make this procedure desirable especially for local area analysis if it is practical and circumstances permit. The forecaster must focus his attention on many details he might otherwise overlook. Accurate local area forecasting is dependent upon the number of factors considered in the forecast, including such factors as geographic influence, local upper air peculiarities, etc., which the forecaster is capable of imposing upon the large scale flow.

Computer products present an excellent means of determining the macroscale meteorological features. They also fill in the gaps in data scarce areas (oceanic, desert, mountain, etc.). It must be remembered however that computer products also have their weaknesses and that their validity should be verified to the degree possible and practical.

Numerical Prognoses Techniques

The first numerical prognoses techniques used either a "Barotropic Model," FNWC Monterey, or a "Baroclinic Model," National Weather Service. With the use of these techniques the machine product equalled or exceeded in accuracy the product produced by skilled forecasters. However, the early model still missed strong development frequently because of the limited number of meteorological variables considered in the prognoses.

A new model has been developed which uses Newtonian Equations and is called the Primitive Equation (PE) model. The Primitive Equations used in the model are:

1. Newton's Second Law of Motion. The individual motion of a particle of air is the result of the sum of all the forces acting on it.
2. The Thermodynamic Equation. Changes in the potential temperature of a particle are the results of heating and cooling.
3. The Continuity Equation. Mass is conserved.
4. The Conservation of Water Vapor Equation.
5. The Equation of State. Pressure, density, and temperature of a perfect gas are related.
6. The Hydrostatic assumption is used in place of the vertical momentum equation.

The PE-Model vertically divides the Northern Hemisphere atmosphere into 5 layers, six sigma surfaces. (The sigma surfaces are at the bottom, between layers, and at the top of the atmosphere). The basic inputs for the model are:

1. Virtual temperature analyses for the Northern Hemisphere at 12 constant pressure levels distributed from 1000 mbs to 50 mbs.
2. Height analyses at seven constant pressure levels.
3. Moisture analyses at 4 levels from the surface to 500 mbs.
4. Terrain field. (Height of mountains).
5. Sea level pressure analyses.
6. Sea surface temperature analyses.
7. Albedo field, which is computed using the monthly mean surface temperatures.
8. Nondivergent wind components, obtained from the solution of linear balance equations at pressure surfaces with interpolation to sigma surfaces.

During the calculation to solve the PE Model a number of items are considered such as:

1. Surface friction in the lowest layer.
2. Radiation processes, both solar and terrestrial.
3. Sensible heat exchange and evaporation.
4. Condensation and the release of the latent heat of condensation.
5. Moist convective processes.

The results of the solution of the PE Model are surface and upper air prognoses for the northern hemisphere for times out to 72 hours, as well as additional information such as:

1. Large scale precipitation prognoses.
2. Sea and swell height prognoses.

The verification of the PE Model prognoses shows a definite improvement in the FNWC Monterey's product.

1. At the surface the verification shows:

a. During the winter the PE was about twice as skillful as the thickness-advection model.

b. During the summer the improvement was less; however, there are fewer baroclinic situations in summer.

2. At the 500 mb level the results of the verification are excellent. The model also shows good results with its ability to simulate the growth of baroclinic waves.

SURFACE PRESSURE ANALYSIS.–The surface pressure analysis is accomplished through the following program input:

1. Reported surface observations (approximately 4000-5000 reports).

2. Hourly history tape of surface pressure prognosis derived from the preceding 0000Z or 1200Z analysis.

3. Hourly history tape of 500 mb prognosis derived from the preceding 0000Z or 1200Z analysis.

4. Surface pressure climatology field for the present month.

5. Hemispheric surface coverage for the preceding 0000Z or 1200Z analysis.

PROGRAM COMPUTATIONS.–Construction of the "first guess" is as follows: The first guess to the surface pressure analysis is a modified surface pressure prognosis which verifies at map time. This map is a result of combining the 6-hour old surface analysis and 3 hour surface prognosis verifying 3 hours prior to map time for the procedure of updating and reanalysis. A description of the procedures used in reanalysis is presented in chapter 3 of the Computer Products Manual, NavAir 50-1G-522.

Program Limitations

To intelligently utilize the computer products derived from the computer program it is important that the forecaster understand the program limitations as well as its advantages. Some of these limitations are presented in the following paragraphs:

1. The surface pressure analysis makes use of reported pressures and ship winds only. No attempt is made to model the pressure distribution using other elements of the reports. For example, isobars crossing a front will be kinked only where reports are sufficiently dense to so indicate; a heavy rain report will not be considered in the analysis of a frontal wave.

2. Analysis flaws are usually the result of one or all of the following problems:

a. Lack of data.

b. Erroneous data.

c. Extreme atmospheric change.

3. In regions of no reported data, the final analysis will be a combination of the first approximation and climatology. Initially the contribution made by climatological values is small but increases with the number of synoptic periods in which no reports are received in the area.

4. Surface observations reported in error may be used in the analysis. No internal consistency checking of reports is presently attempted. A ship reporting ten degrees out of position or ten millibars off the actual pressure due to transmission error, for example, will not be corrected. The data will either pass the error checks, resulting in an incorrect analysis, or it will be rejected with the possible loss of some significant information.

5. Surface pressure and ship wind observations reported in error will normally be rejected in the Gross Error Check (GEC). If the erroneous report should pass the GEC tolerances, it will contaminate the analysis.

6. If the bad report, accepted in the GEC, is located in a high data density region (e.g., over land), it will probably be rejected when compared with its neighbors in the lateral fitting check. It may, however, have already had a slight adverse influence on the final analysis.

7. If the bad report is located in a sparse region (e.g., over oceans), it may pass the lateral fitting check as well as the Gross Error Check. In this case the analysis will depict with utmost precision the pressure distribution represented by the erroneous data.

8. Occasionally extreme atmospheric change will manifest itself as an error in the analysis. In this case, strong development results in valid data failing the gross error check tolerance. Gross error check tolerances are predicated on a reasonable forecast being used in construction of

the first approximation. If the forecast is abnormally poor and off-time and late reports do not correct it sufficiently, this type of error may occur. Although rare, it represents a serious flaw and should be guarded against.

Detection of Errors

Errors of the types previously discussed can easily be discerned at major network centers. Data density upon which the analysis is based is provided by the Hemispheric Surface Coverage and Gross Error Check Reject Coverage Charts. Data lists, including all ship reports used in the analysis as well as those rejected by the Gross Error and lateral fitting checks are available. It is expected that major network centers will use this information to determine validity of the surface pressure analysis in their areas of responsibility. Corrections, if necessary, will be initiated or notices of error will be disseminated to fleet users, by these centers.

Errors may be detected by the operational forecaster who receives the analysis directly on the high-speed communication links. The total number of reports used, a simple indicator of analysis quality, appears in the title of all surface pressure analysis. The 24 hour surface pressure change chart or message can be used to determine areas of abnormal surface development. Nonmeteorological or new disturbances should be cross checked with available observations.

Verification

A quick method of verifying the computer analysis is to take the hand drawn analysis having the same time and lay one over the other on a light table. More complex methods involving the use of acetate overlays, ozaiid machines, and overhead projectors, are sometimes employed at larger weather units.

Weather units may receive computer products via facsimile or teletype message. The facsimile may be difficult to verify using the light table method due to variations in transparency. In this case it may be necessary to verify by replotting coordinates from one chart to the other to depict the significant chart features. Teletype messages in code form are also transcribed onto appropriate charts following the coordinate plotting procedure.

Isobaric Consistency

Once the validity of the computer product has been determined the isobaric patterns of the various charts should be compared for uniformity. The adjustments made between charts will involve variables such as the forecaster's experience, chart scales, operational objectives, etc. In any case, the computer products should be compared prior to finalizing the hand analysis. The types of computer products available and transmission times are presented in various publications and directives including H.O. 118; Computer Products Manual, NavAir 50-1G-522; NMS Forecasters Handbook No. 1.

ADDITIONAL RULES AND CONSIDERATIONS

Mountainous terrain presents a distinct and difficult problem in surface analysis because of the roughness of terrain and the uncertainty in the reduction of pressures to sea level. In this case some higher level, either 850- or 700-mb, should be consulted as a guide to the reality of pressure centers and to obtain reasonable correspondence between the pressure patterns as represented on the surface map. Along the mountains you may have a packing of the isobars with an unrealistic gradient and winds may even blow at right angles to the isobars. You should become familiar with the local peculiarities of mountainous terrain and make allowances for them when analyzing the surface chart.

Along coastlines, especially the Atlantic and Gulf Coasts of the United States during winter, a problem of isobaric analysis often arises when cold air of polar or Arctic origin flows over much warmer coastal water. Since it is found by observation that a relatively smooth flow pattern continues aloft (850- or 700-mb and higher), it follows that the configuration of sea level isobars at coastlines should be affected markedly by the warming of air in lower levels, because the warming produces lower pressures at the surface over the water than otherwise would be the case. This type of distortion is illustrated in figure 6-6. The dashed lines represent a smooth extrapolation of the pattern over the ocean, and the solid lines show the true pattern. The magnitude of this effect varies with the temperature difference between land and sea, and presumably with the lapse rate in the cold air when over land. This effect can be anticipated off any coast during winter.

Highly irregular isobars are not impossible but they are improbable. It is a common error for beginners to draw isobars with many irregularities in order to fit them precisely to the pressures and the map. Consequently, the isobars have a wavy appearance which is not corroborated by the wind observations. Errors in barometer reading may be responsible for the difficulties in fitting the isobars to the wind system. At sea, where readings from aneroid barometers are not highly accurate, there is a tendency for inexperienced analysts to draw wavy isobars on maps. While you should make an attempt to smooth out the lines on the chart, you should not try to smooth out irregularities which are judged to be real and convey important information. At sea, the effect of friction is

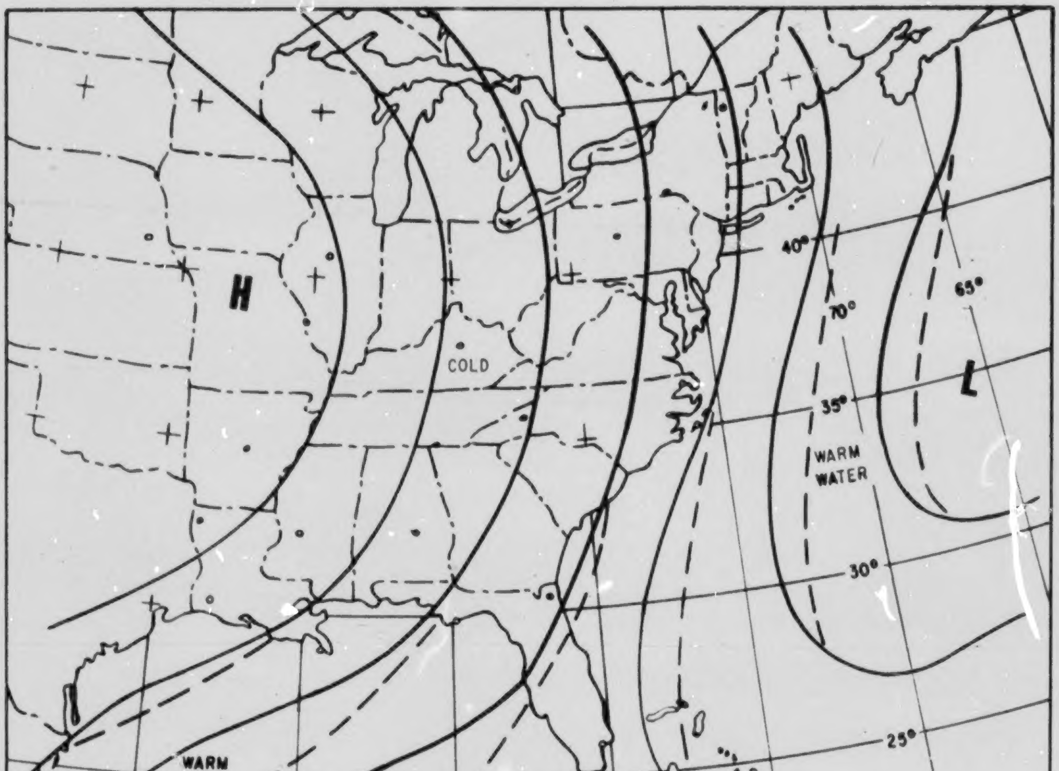

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Figure 6-6.—Effect of warm coastal waters on isobars.

minimized and the wind observations are largely dependable. Pressures are not as accurate, so more weight should be given to winds than pressures.

SUMMARY OF RULES FOR DRAWING ISOBARS

The order given is not arbitrary and may be varied to suit local needs or procedures. This summary includes some of the suggestions for drawing isobars on surface charts.

1. Choose the isobar interval according to the previous instructions.
2. Select isobars which will outline pressure centers best and draw these centers first, using history and winds to help locate them.
3. Sketch lightly the orientation and spacing of isobars in the vicinity of isolated reports; elsewhere they may be sketched in a little heavier.
4. Draw isobars at largest interval commensurate with ability and experience. Inexperienced analysts will find the 8-mb interval about the limit. Experienced analysts are able to use 16-mb intervals in their first approximation of the isobaric pattern. Intermediate 8-mb isobars are then added, followed by 4-mb isobars. Erasures and adjustments to the initial patterns are usually necessary and to be expected in the final analysis.
5. Along an axis joining two highs or two lows, two isobars with the same label must occur; along an axis joining a high and a low, two isobars cannot have the same label.
6. Serious errors in circulation patterns can often be avoided by drawing isobars as if the pencil were traveling with the wind, counterclockwise around lows and clockwise around highs. This will help prevent placing a small closed center on the wrong side of an isolated wind report.
7. Resolve conflicts between wind and pressure data and smooth isobars by evaluating the reliability and representativeness of the data.

FRONTAL ANALYSIS

When a front is drawn on the weather map the analyst considers many factors which may not be apparent on the map. Location of the various observation stations and local effects play a great part in weighing the elements which are used in locating fronts. The front may be weak, it may be strong, or it may be of moderate intensity. It may be well defined in the isobaric pattern and easy to locate through the surface weather changes peculiar to that type of front. On the other hand, the front may be weak or indistinct and therefore difficult to locate from surface parameters. Consideration of frontal structures from upper air information is helpful in locating all types of fronts. These features of fronts on upper air charts are discussed in a later section.

When a front approaches and passes a land station, the sequence of events may be anticipated, but at sea the movement of the ship combines with the movement of the front to produce changes which are not so easy to foresee. For example, on a ship traveling westward the front passes quickly, for the front and the ship are moving in opposite directions. If the ship is moving eastward, in about the same direction as the front, the change of wind and attendant weather conditions take place slowly. In exceptional cases a fast ship may overtake a front and pass through it from west to east, thus reversing the sequence of changes. Variations in the type of weather encountered with frontal passages must also be taken into account. The typical models available only represent typical conditions and other factors such as strength of the front and nature of the terrain over which it has passed or is passing, and subsequent modifications to the air masses, speed of the front, and type of front must be taken into consideration. In this section of the chapter, some of the rules and aids in locating fronts on the surface weather map are discussed.

LOCATION OF FRONT ON A WEATHER MAP

After the data has been entered and its reliability evaluated on the map, one of the major tasks of the analyst is locating the fronts. Some of the factors which enter in the determination of fronts on a weather chart are as follows:

1. Fronts should show logical continuity from previous charts. Corrections should be taken into account.

2. Fronts normally lie in troughs of low pressure. While a trough of low pressure in the pressure pattern is a necessary condition for a front, it is not a sufficient condition. The front must move in a manner consistent with the motion of the colder of the two air masses. In other words, a frontal surface is composed of particles of air which move with the winds. If the trough line does not move with approximately the component of the winds normal to it, the trough line cannot contain a front. This is not, however, sufficient by itself, for a trough line could move at the proper speed and still not contain a front. Other factors must be considered.

3. A moving front will have a pressure tendency difference across the front. If the front is stationary or slow moving, there will be little or no tendency difference across the front. If the front has passed the station within 3 hours of map time the pressure tendency will show a kink.

4. Winds will shift cyclonically across the front.

5. Dewpoint differences will exist across the front. Precipitation and water surfaces affect dewpoint values so, at times, they are not representative of air masses.

6. Temperature, at times, is a good indicator for locating a front; however, the temperature of the air near the ground is so dependent upon insolation and radiation that it is often not representative of the air masses. If two stations at nearly the same elevation are cloudy and have strong winds, then a temperature difference may not be significant.

7. The clouds and precipitation patterns should be representative of the particular front in question.

8. Fronts should move with the cold air winds and can be located by historical sequence. If the cold air is moving toward the front it will be a cold front; if the cold air is moving away from the front it will be a warm front.

9. The line representing a front should be placed on the warm air side of the transition zone and along a line of cyclonic wind shear.

In summary, in the analysis of a weather map for fronts, the foregoing steps should be considered together. Along a single front the data used to locate a front may vary considerably. For example, at one point a temperature difference may give the strongest indication, while wind may be locally influenced and not give a representative picture of the frontal location. Other parts of the fronts may be more easily identified by pressure tendency or wind. No one item should be used as a complete guide to analysis.

ISOBARS IN RELATION TO THE LOCATION OF FRONTS

The isobaric pattern reflecting the traditional kink of isobars at fronts is often one of the best indicators of the existence of a front on the weather chart. The amount of kink is governed by the strength of the front, its movement, and the particular stage of development.

With due consideration to past history and the other factors mentioned above, a good method, especially at sea, is that of drawing down the wind along observations where a front is suspected to exist. For example, in figure 6-7, suppose you start to draw the isobar for the 08.0 (1,008 millibars). Drawing down the wind in proper relation to the barometer readings, with the wind at small angles (aboard ship) across the isobars toward low pressure, you come to observation point B. Here you will note that the next observation ahead shows a change in the wind, and it is now necessary to draw the isobar toward C. It is evident that there is a discontinuity, or front, along the dashed line from D to E. If you draw all the isobars carefully, studying the winds as you go; the isobars would look like those in figure 6-8.

This shows you that there are irregularities or discontinuities which are not a result of error in observations, for they are systemically arranged. The best fit of the isobars is obtained by making an angle in the isobar with the vertex at the front, or wind shift line.

Figure 6-8 shows the location of fronts in a cyclonic system approaching the west coast. The fronts in this situation are clearly identified by the reports. Fronts should be marked on the map when they can be clearly identified.

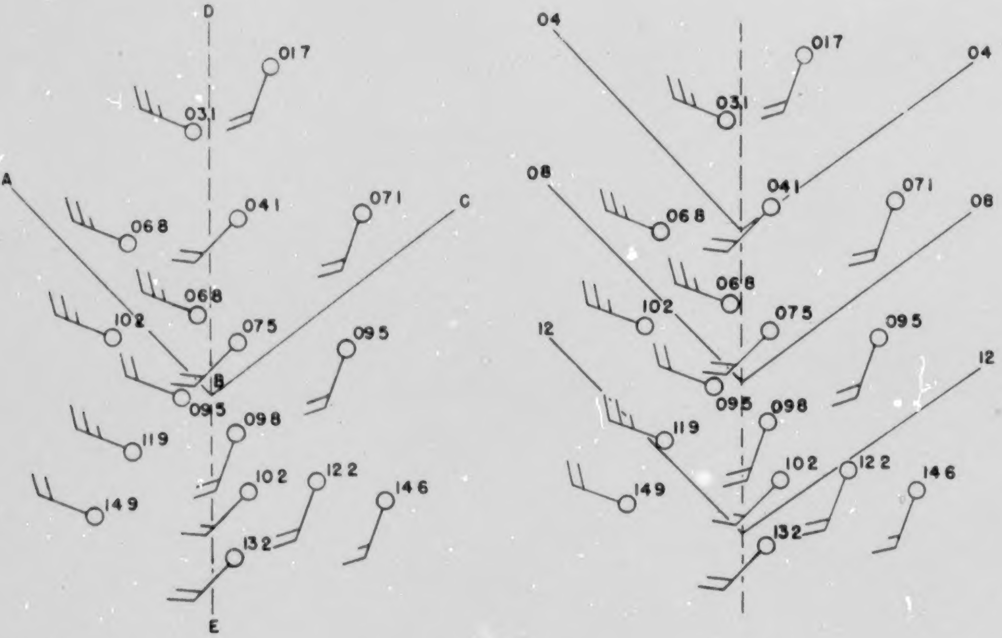

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Figure 6-7.—Isobaric discontinuity at a front. At left, drawing down the wind from A, a discontinuity is found along line DE. At right, the isobars show a systematic arrangement.

CONSIDERATION OF ELEMENTS

Assume that a cold front is to be located. Local analysis may reveal that the pressure trough, the wind shift, the line of showers, and discontinuities in temperature, moisture, and tendency coincide at most in pairs. What line should be chosen to represent the front? The following procedure is recommended:

1. Eliminate erroneous and unrepresentative data in accordance with previous instructions in this chapter.
2. Locate the front at selected points where the maximum number of identifying characteristics occur simultaneously.
3. In the absence of known orographic obstructions, interpolate or extrapolate the front in doubtful or confused areas.
4. Check the 3-dimensional geometrical and meteorological consistency of the frontal surface at other levels.
5. Check the displacement of the frontal surface from its last known position for consistency with movement of the colder air mass during the interval.

Step 5 is by far the most important and should be applied twice, once in the beginning to determine approximately where the cold front begins and again at the end for a final check.

It is convenient to delineate only the leading edge of a cold frontal zone and the trailing edge of a warm frontal zone with their respective frontal symbols. This puts all the strong gradient of temperature in the cold air mass.

INTERPRETATION OF SATELLITE CLOUD PHOTOGRAPHS

Satellite cloud photographs have proven to be a most valuable aid in the location of fronts on the weather map. It is important for the analyst

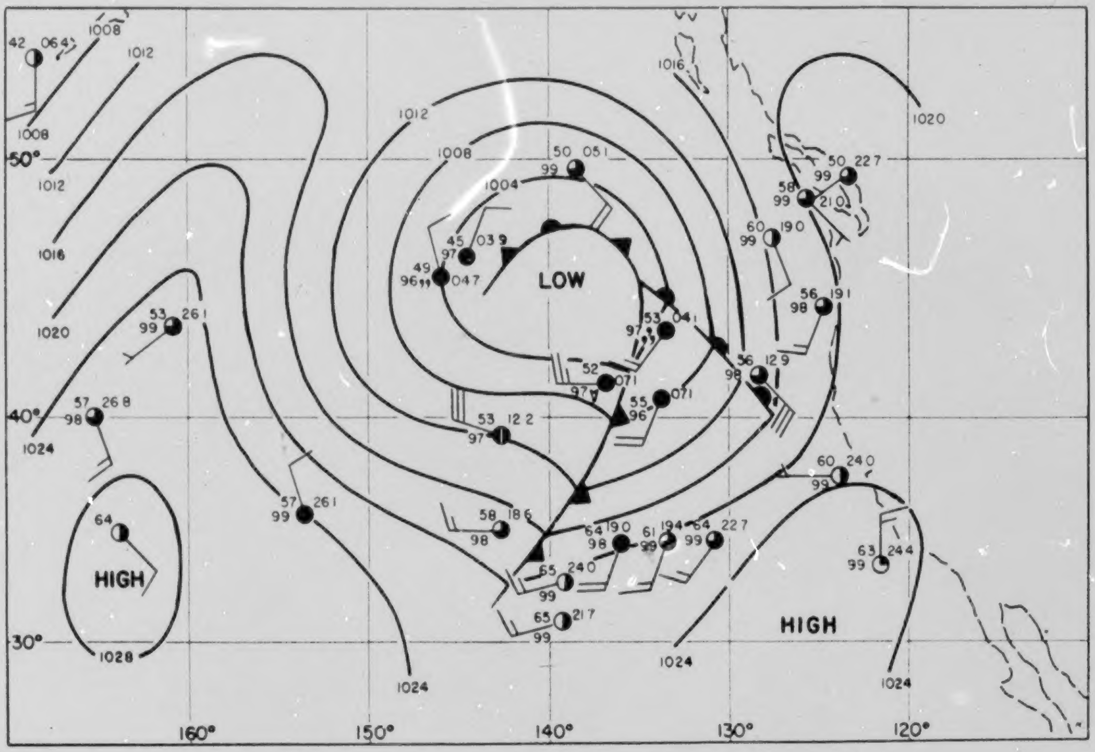

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Figure 6-8.—Surface weather map showing location of pressure centers and fronts from isobars.

to keep in mind, however, that the satellite is presenting a view from the top down. This means that features which appear only in the lower layers of the atmosphere might be hidden by extensive cloud coverage aloft. While satellite pictures are undoubtedly a tremendous step forward for the analyst, some analysts fall into the trap of blindly accepting them as infallible. The satellite photograph must be viewed in proper context, with as many other pertinent factors as available being added to the final analysis. It cannot be overemphasized that the analyst must use all of the tools at his disposal if he is to acquire consistent accuracy in his analysis.

Identifying Cloud Systems

An example of the appearance of frontal cloud structure in satellite cloud pictures as compared with surface analyses is shown in figures 6-9 and 6-10.

These two figures illustrate a sequence of events over a 24 hour period. Notice the similarity in the configuration of the clouds and fronts depicted in the analyses and photographs. The center of circulation is outlined in the photographs by the entrance of the cold dry air in the upper levels during the occlusion process. Notice the breaking down of the organization in the cloud structure as the occlusion or dissipating process progresses during the time period represented by the two figures. The utility of these photographs in identifying the cloud bands along the periphery of the storm area is quite evident. This is especially true in sparse data areas.

The analyst must bear in mind the normal slope of fronts and pressure systems aloft when

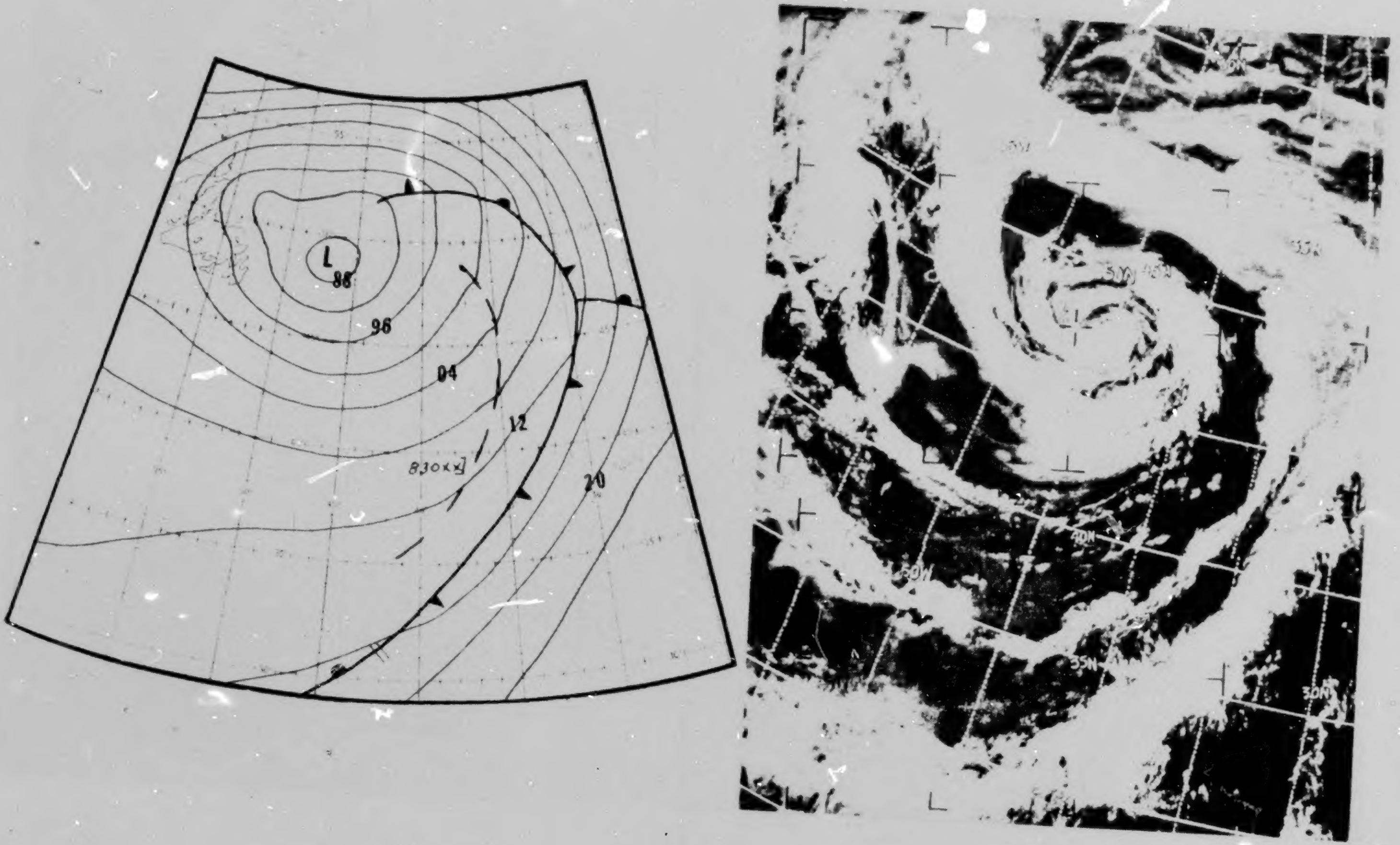
L
88
96
04
12
20
830xx
50N
45N
40N
35N
30N

Figure 6-9.—Comparison of surface analysis and cloud pictures on 31 May 1969.

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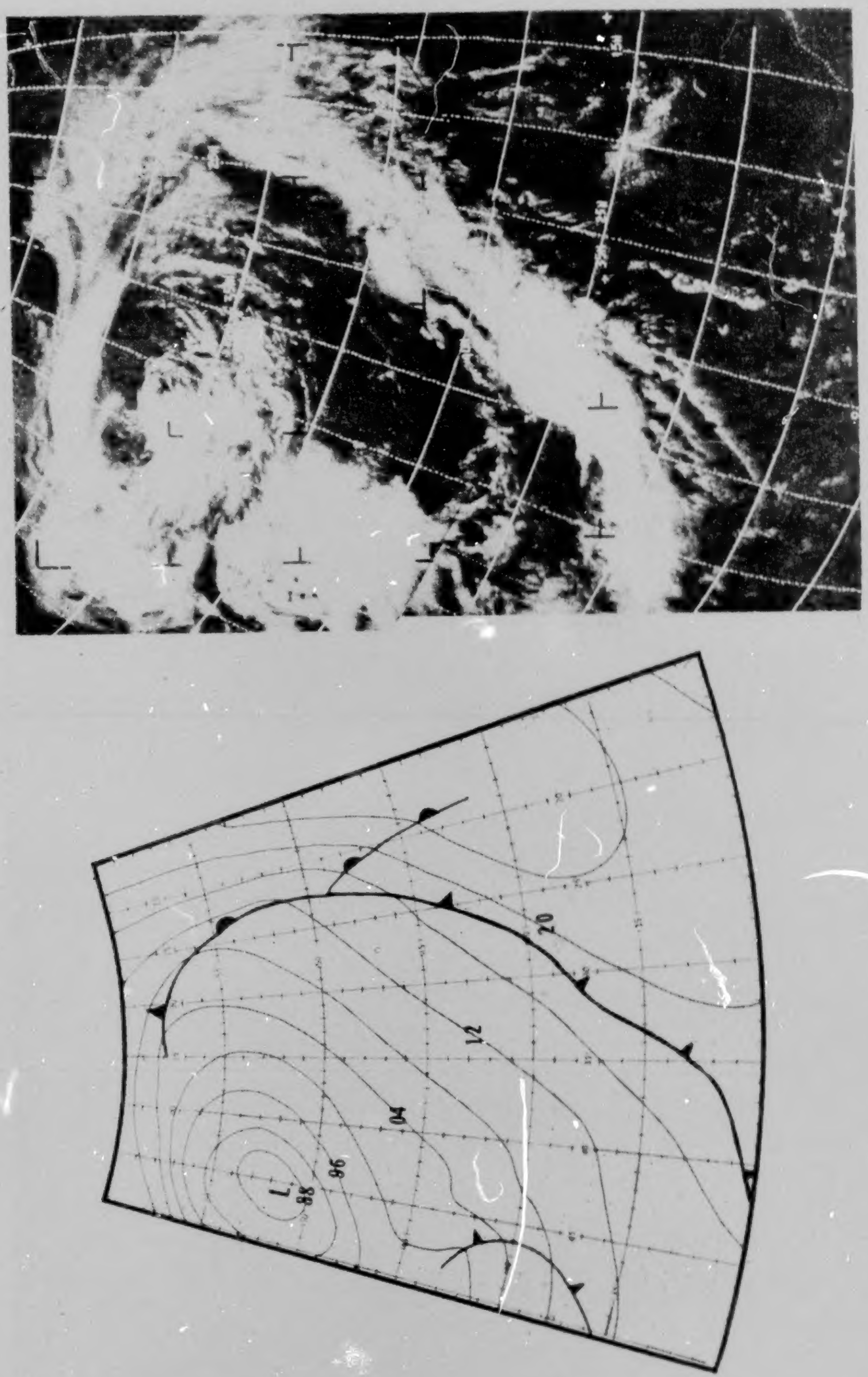
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Figure 6-10.–Comparison of surface analysis and cloud pictures on 1 June 1969.

making comparisons between surface analysis and the satellite pictures. More discussion related to this subject is presented in the chapter following this one pertaining to upper air analysis as well as in chapters 8 through 12 of this manual pertaining to prognosis. Discussion of the location of surface frontal positions on the satellite photograph may seem somewhat premature at this point in the manual due to the considerations which must be given to upper air systems. However, their usefulness in estimating frontal positions must be considered from the outset of the analysis.

More detailed information pertaining to the utilization of satellite photographs in analysis and forecasting may be found in the Direct Transmission System Users Guide published by the National Oceanic and Atmospheric Administration (NOAA); the Guide For Observing the Environment With Satellite Infrared Imagery, NWRF F-0970-158; chapter 8 of the NMS Forecasters Handbook No. 1, Facsimile Products; and other NWRF booklets as listed in Navy Weather Research Facility Reports and Publications, NWRF 00-0369-143.

Direct Readout Infrared (DRIR) Interpretation

The foregoing discussion has been with reference to the commonly received camera sensed satellite photographs. However, the infrared radiometer is now frequently used as a satellite sensor and the analyst or forecaster must endeavor to properly interpret the information received.

An infrared sensor differs from a visual (camera type) sensor in that the instrument measures radiated heat rather than reflected light. Infrared sensors therefore measure the temperature of cloud tops (or the earth's surface) rather than their albedo. The infrared display is arranged so that warm bodies (earth, low clouds) appear dark, while cold bodies (high clouds) appear bright. Thus, infrared displays of clouds have an appearance that is similar to visual pictures of clouds, but have an entirely different meaning.

SET ADJUSTMENTS FOR DRIR.—Current satellite radiometers display a 7-step calibration wedge along with the DRIR picture. A calibration table is provided for each radiometer, allowing the analyst to assign a temperature value to each shade of gray in the step wedge. This allows the analyst to interpret the infrared display in terms of temperature. A forecaster should consider operational goals and local conditions to enhance the information that can be obtained from a DRIR display.

The typical radiometer has a temperature range extending from approximately 180°K (-93°C) to 320°K (+47°C). In general, the coldest expected cloud temperatures are reached at the tropopause (approximately 220°K (-53°C) in mid latitudes). The warmest temperature of interest will be about 293°K (20°C). Therefore, for meteorological purposes, the entire range of the sensor is not required. APT sets should be adjusted so that all step wedges for temperatures colder than the tropopause temperature (220°K) appear white. Similarly, all step wedges warmer than a nominal surface air temperature (293°K) should be adjusted to appear black. With this adjustment, cloud features will appear in finer detail.

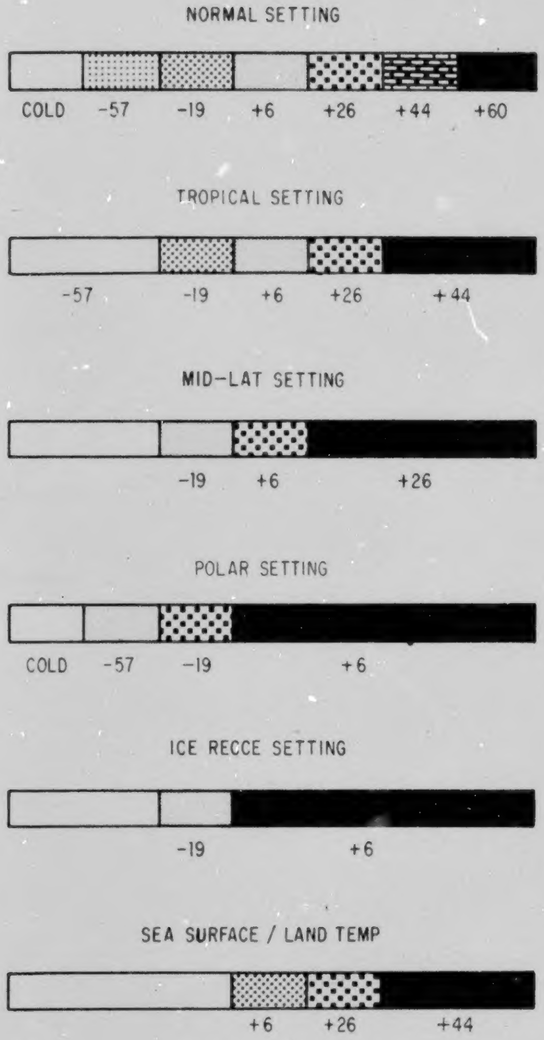
In a similar fashion, the DRIR-scan step wedges can be adjusted to highlight other features of interest, based on their temperature characteristics, as shown in figure 6-11.

DISTINCTIVE DRIR FEATURES.—In terms of the earth and its atmosphere, on a nominally adjusted readout set the various features should appear as follows:

1. Hot deserts should appear black.
2. Lakes and oceans should appear near black.
3. Clouds with low tops (Cu, Sc, St) should appear dark gray.
4. Clouds with medium tops (As, Sc) should appear light gray.
5. Clouds with high tops (Cb, Ci, Cu) should appear white.
6. Snow and ice should appear light gray.

Because of the nature of infrared sensing, however, an operator who understands infrared display can adjust his readout equipment to enhance features of interest.

Figure 6-12 illustrates two typical infrared passes from Nimbus IV over the Western United States and Mexico.



AG.506

Figure 6-11.—Idealized display of DRIR step wedges with readout calibrated for specific interpretation tasks. Numbers below gray shades are equivalent scene temperatures in degrees Celsius.

Figure 6-12(A) is a daytime view, while figure 6-12(B) is the nighttime view of the same area. Note the desert area of Baja California. In the afternoon (figure 6-12(A)), the desert is hot and looks much darker than the surrounding water. In the early morning (figure 6-12(B)), the land has cooled so much that it is now slightly colder, and therefore lighter than the water.

FRONTAL WAVES AND DEVELOPING CYCLONES.—A surface wave, which is indicated in the visual mode by a broadening of the frontal band, has a similar appearance in the IR mode, but with important differences. Cold temperature radiation from the cores of large convective clouds appears as globs of intense white. While the wave cloudiness usually has a uniform white appearance in video pictures, the IR presentation exhibits several shades of gray, indicating clouds of varying height and thickness.

Frontal zones contain stable cloud forms with low cloud tops, as shown in figure 6-13(A), or unstable cloud forms with high cloud tops, as shown in figure 6-13(B).

Since the stability of the air masses in a frontal zone may be inferred from the height of the clouds, it follows that stability may be inferred from gray shades in the IR display.

Figure 6-14 illustrates the differences between the stable and active frontal zones as normally seen in a DRIR display.

Stable frontal bands (fig. 6-14(A)) appear gray (warm) with isolated cloud buildups (white) along the band. Active frontal bands (fig. 6-14(B)) appear offwhite, with lines of convective activity (white) within the band.

The inability to assess stability from visual cloud pictures can result in falsely identifying cloud bands as frontal bands, while in fact they may be only associated with low-level eddies. A classic example of this is seen in the low-level vortex shown in figure 6-15.

In this IR scan two approximately circular gray patterns appear near A and B and a comma-shaped pattern at C. Vortex centers are barely visible at A and B. These two circular gray areas represent cloud vortices made up of spiraling cumuliform cloud lines. These do not represent frontal characteristics and are shown as lows on the surface analysis. With intensification these systems could become frontal. Notice

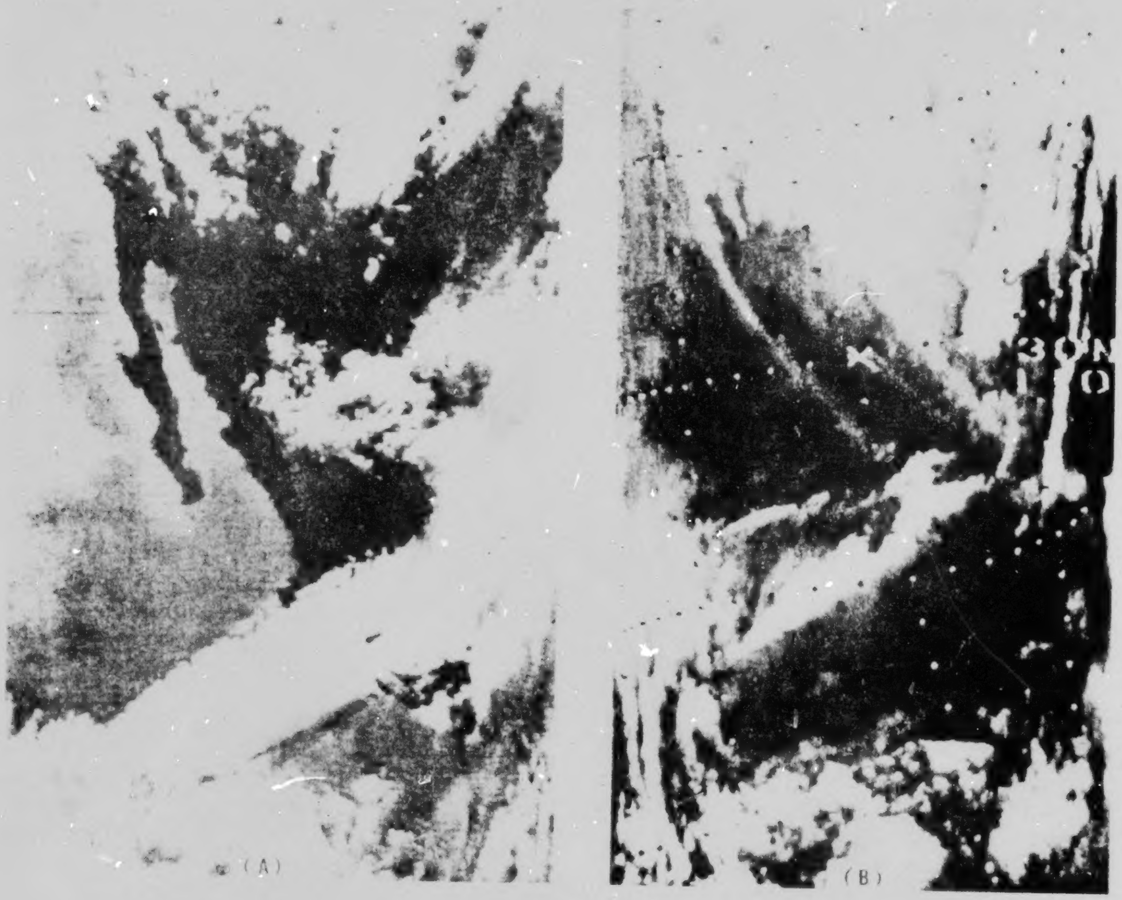


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Figure 6-12.—DRIR picture. (A) Nimbus IV daytime infrared view of Baja, California; (B) Nimbus IV nighttime infrared view of Baja, California.

that this is a nighttime readout. Due to the increased temperature contrast in the lower levels, nighttime readout doubles the opportunity for early discovery of new disturbances, and provides continuity on existing ones.

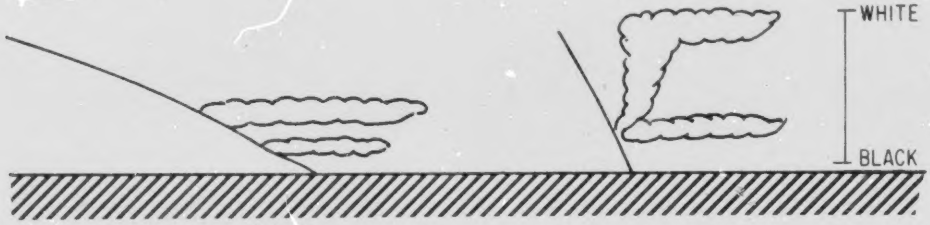
A frontal wave which appears gray in the infrared can be expected to be stable and unlikely to develop, whereas an area of extensive white cloudiness (cold temperatures) indicates that the wave is unstable and is in the process of development. These conditions are further illustrated in figure 6-16.

Some characteristics of the IR mode which favor wave development are:

1. Evidence of a significant increase in the area of cold temperatures (high clouds).

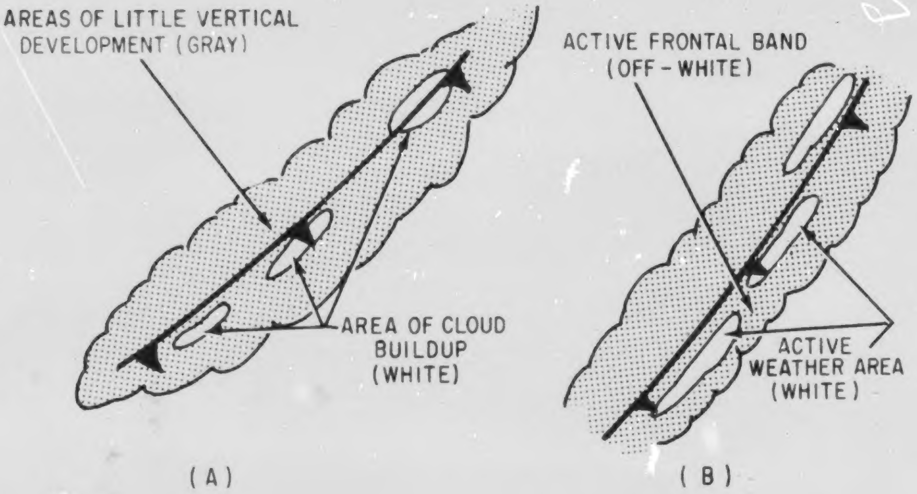
2. Evidence of a jet stream near the poleward boundary of a bulge on the frontal band.

3. Positive Vorticity Advection (PVA MAX) or strong vorticity center, such as those described in chapter 7 of this manual, upstream from the frontal wave. This would indicate the presence of a short-wave 500-mb trough upstream from the developing cyclone.


WHITE
BLACK

AG.508

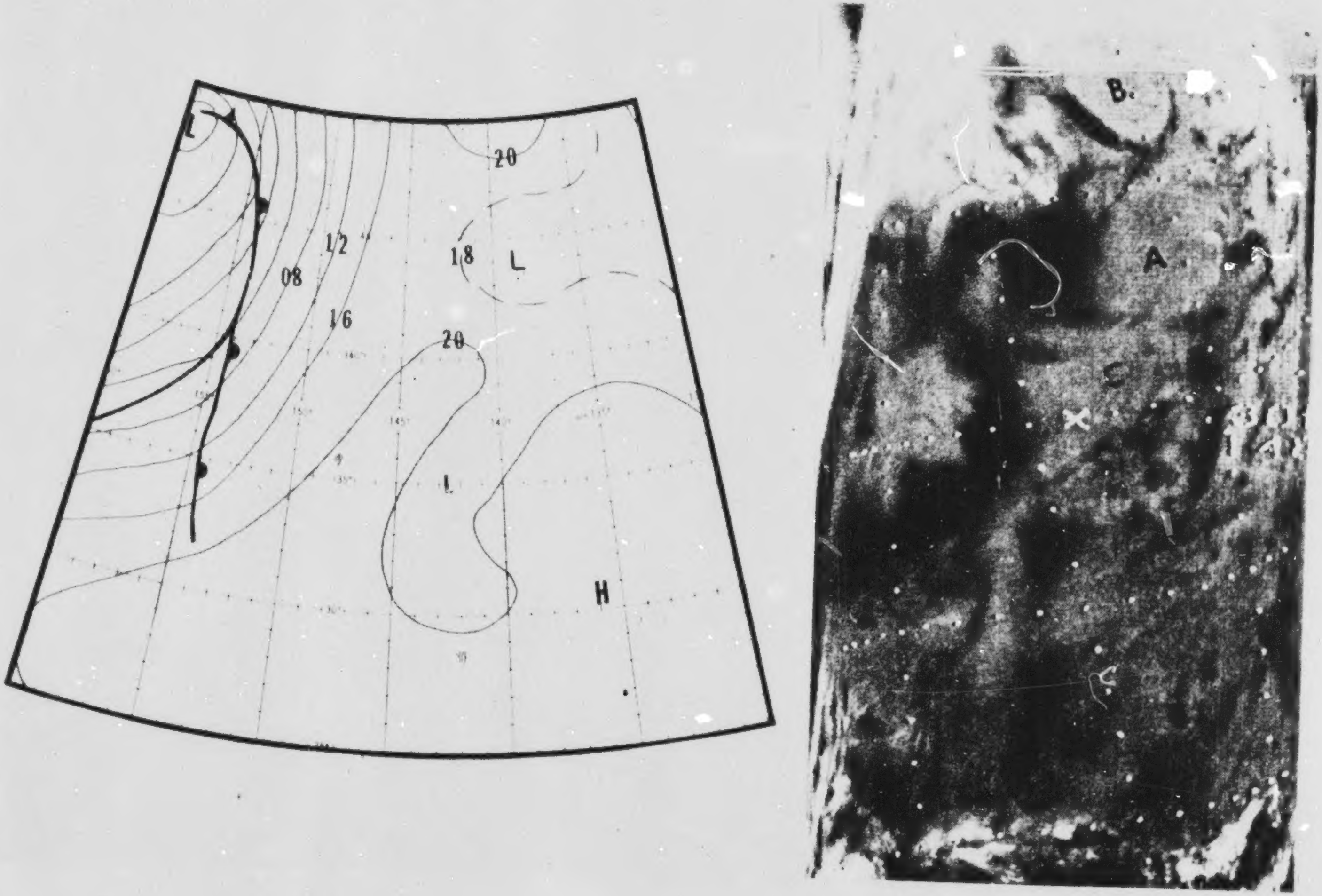
Figure 6-13.—DRIR cloud depiction. (A) Stable cloud form; (B) unstable cloud form.


AREAS OF LITTLE VERTICAL DEVELOPMENT (GRAY)
AREA OF CLOUD BUILDUP (WHITE)
(A)
ACTIVE FRONTAL BAND (OFF-WHITE)
ACTIVE WEATHER AREA (WHITE)
(B)

AG.509

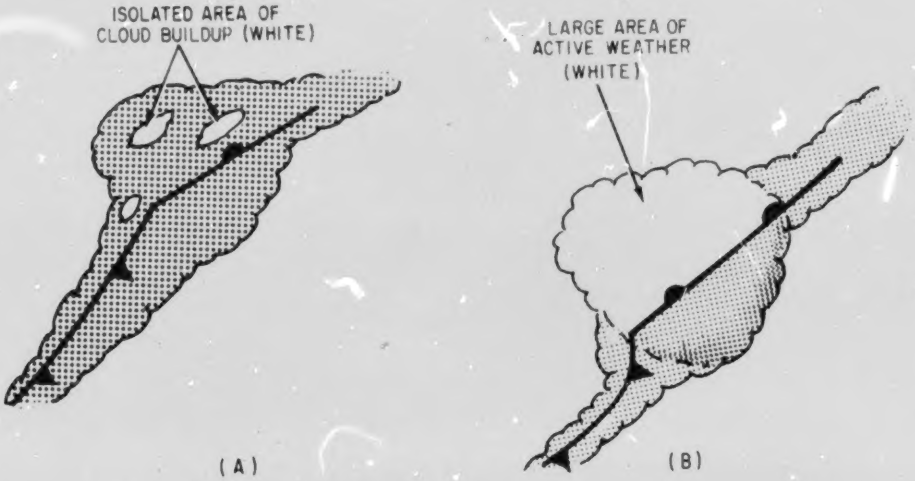
Figure 6-14—DRIR frontal depiction. (A) Inactive frontal zone; (B) active frontal zone.

When an unstable wave (cloud bulge) is developing into a vortex, (figure 6-17) the first evidence is the beginning of a dry tongue on the trailing edge of the cloud band. The cells (open) normally seen in the dry tongue over the ocean during the Northern Hemisphere summer appear mottled gray in the IR mode. Cold air flowing over warm water creates low-level instability and convective overturning. The greater the air/sea temperature difference, the greater the degree of overturning. In areas of cold air advection (dry tongue), low cumulus cloud tops (gray) indicate stable conditions and a small air/sea temperature difference. High cumulus cloud tops (white) suggest a large air/sea temperature difference and strong cold air advection. Due to the higher cloud tops of the cumulus congestus, cells (open) have a similar appearance in the IR and visual modes. If the field of cells (open) is bright in the IR mode, this indicates intense cold air


L
20
12
08
16
18
L
20
L
H
B.
A
C
X
C

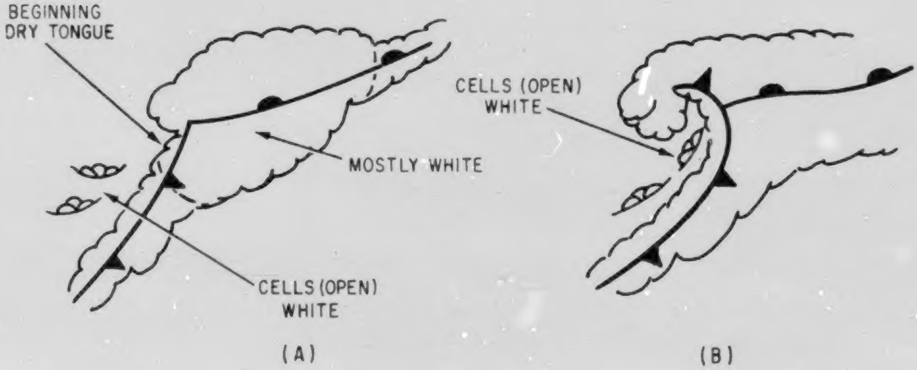
AG.510

Figure 6-15.—Surface analysis and accompanying DRIR picture showing low level vortices.


ISOLATED AREA OF
CLOUD BUILDUP (WHITE)
LARGE AREA OF
ACTIVE WEATHER
(WHITE)
(A)
(B)

AG.511

Figure 6-16.–DRIR frontal depiction. (A) Stable frontal wave; (B) unstable frontal wave.


BEGINNING
DRY TONGUE
MOSTLY WHITE
CELLS (OPEN)
WHITE
CELLS (OPEN)
WHITE
(A)
(B)

AG.512

Figure 6-17.–DRIR frontal depiction. (A) Developing frontal wave; (B) occluding cyclone.

injection into the cyclone, and increased likelihood of deepening.

Additional information pertaining to the uses of DRIR readouts in forecasting will be presented in chapter 9 of this training manual.

COMMON ERRORS IN FRONTAL ANALYSIS

Some of the more common errors in frontal analysis are listed below:

1. Inconsistent displacements from previous positions.
2. Cold fronts improperly designated as warm fronts and vice versa.
3. Too many fronts, particularly secondary fronts.
4. Isobars too sharply kinked at fronts or kinked improperly toward low pressure.
5. Frontal patterns in the horizontal which have an impossible 3-dimensional structure.
6. Use of unrepresentative data (particularly temperature) in locating fronts.
7. Dropping of fronts in areas of sparse or no reports without designating frontolysis on preceding chart or charts.

STATIONARY AND NONFRONTAL TROUGHS

Stationary troughs can contain fronts only momentarily. The most common such trough in North America occurs on the lee (east) side of the Continental Divide. Unless the air mass following the front is dense enough to fill the trough, a front moving eastward across the Rocky Mountains may occupy this trough at a particular map time, but will move on through while the trough remains fixed. Needless to say, warm fronts moving through such a trough will not affect its stationary character. These troughs occur on the lee side of any mountain range orientated across a deep current of air. Another kind of stationary trough occurs in southwestern United States, and is most pronounced in summer, the season of the "Heat Low," when it sometimes extends the full length of the west coast.

The most common nonfrontal troughs in middle latitude ocean areas occur in the wake of well developed occluded cyclones. (They also occur over land areas.) This trough is often incorrectly analyzed as a secondary cold front because it has a well defined line of showers, cumuliform clouds, wind changes, and pressure tendencies across the trough line similar to those of the cold front. Nevertheless, it moves at speeds very much less than that of the often very strong winds which accompany any deep cyclone. If this phenomenon occurs in an area with a sufficiently dense network of reports, an analysis of the wind field shows that there is a definite line of longitudinal convergence in the trough, the result of a "surge" in the northwesterly or westerly winds of the polar or arctic air mass moving in behind the occluded cyclone.

SATELLITE DEPICTION OF SURFACE RIDGE LINE

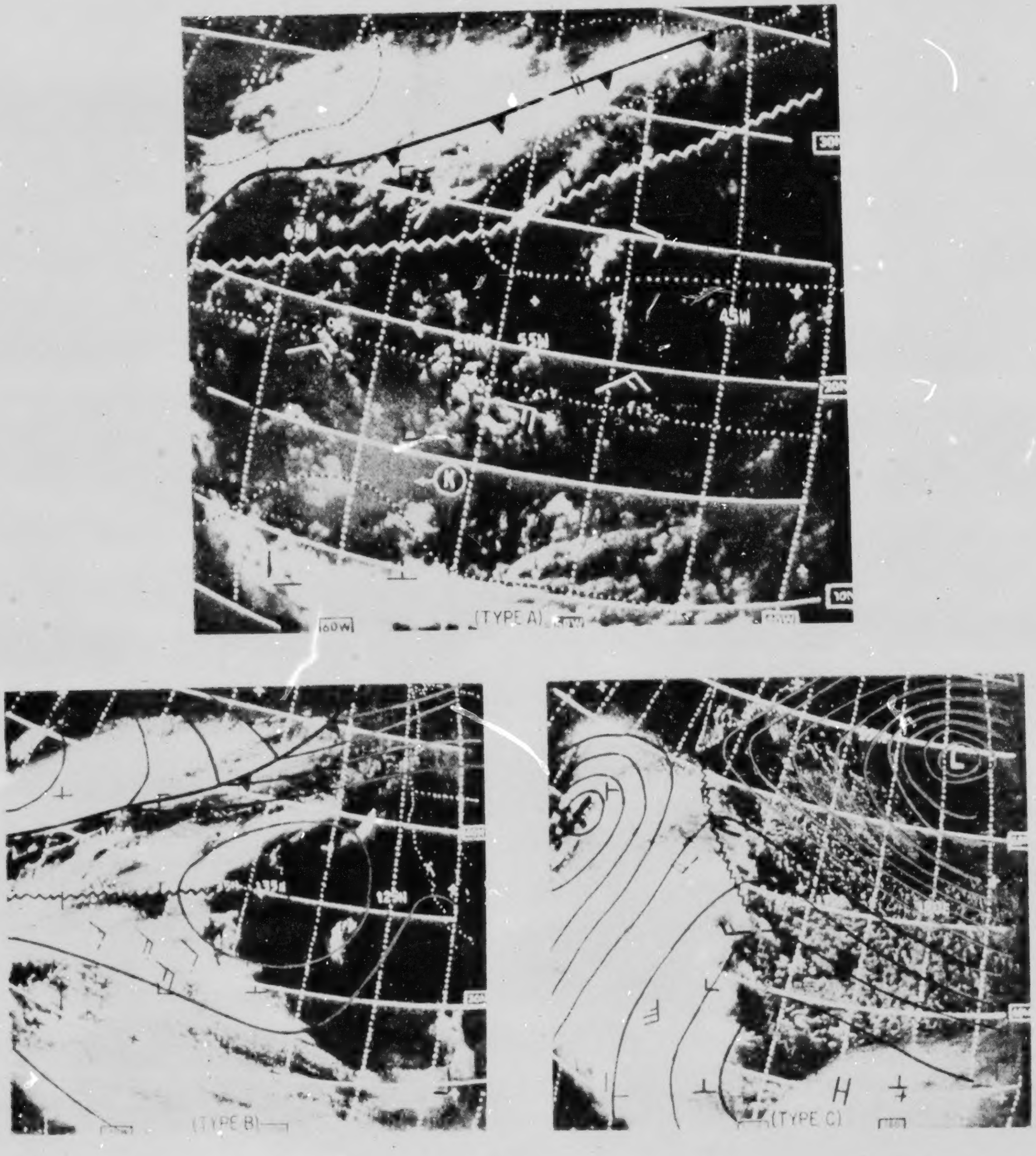
Among the nonfrontal meteorological features which may be confirmed by the satellite photograph is the surface ridge line. Although they may not be as well defined as frontal systems, ridge lines may be identified with increasing accuracy commensurate with the experience of the analyst.

Three characteristic cloud patterns found with surface ridge lines on satellite photographs are illustrated in figure 6-18.

The cloud pattern of type A in figure 6-18 is characterized by long cloud lines or fingers which extend, in a continuous fashion, from a frontal band. These fingers generally are oriented in a more north-south direction than the frontal band. The end of these continuous cloud fingers serve as indicators or points for positioning the surface ridge line on the southwestern side of a subtropical anticyclone.

The type B (fig. 6-18) surface ridge line generally is found on the western side of a subtropical high where the clouds change in character from cumuliform to stratiform. Notice in figure 6-18 that the points drawn for the ridge line are drawn so that they lie along the line where the cloud type changes from cumuliform to stratiform. This change in cloud character is generally found where the low level wind changes from a southeasterly to a southwesterly direction. As the air moves poleward more rapidly, stability increases near the surface and produces the observed change in cloud form.

The type C pattern (figure 6-18) is associated with a sharp migrating surface ridge that lies between two large cyclones in close proximity to each other. Along this ridge, there is a wind shift from a general southwesterly direction to the northwest. While an air parcel is in the southerly flow, it usually is being cooled from below and thus becomes more stable with time. Air that is in the northerly flow normally is heated from below and becomes unstable. This difference in stability produces two distinctly


30N
65W
45W
20N
55W
20N
10N
60W
(TYPE A)
(TYPE B)
H
(TYPE C)

AG.513

Figure 6-18.—Surface analysis superimposed on satellite pictures showing surface ridge lines.

different cloud patterns that can be recognized in satellite pictures. The clouds in the southerly flow are primarily stratiform, while those in the northerly flow are cumuliform. Hence, the surface ridge is located in the narrow region between the stratiform and cumuliform clouds.

SUMMARY OF SURFACE FRONTAL ANALYSIS PROCEDURES

Probable Location of the Front (Extended Analysis)

If one previous location of the front is known, the current position is downwind, (referring to the wind in the cold air) at a distance depending roughly on the speed of the wind in the cold air.

Generally, cold fronts are found to the east and/or southeast of previous positions, warm fronts to the north and/or east of previous positions, and quasi-stationary fronts in about the same position as before.

When the isobars on the previous map are parallel to the front very little, if any, movement is expected.

When two successive previous positions are known, an approximate past rate of frontal motion is established. In this case the quickest method of locating the probable current position is to lay off the previous displacement, using the pencil and finger technique. The precise location is then obtained by minor adjustments based on the criteria below.

Do not make major adjustments in location from what would be expected by previous displacements unless the wind and/or pressure gradient in the cold air has changed a proportional amount.

Criteria for More Precise Location of the Front (Local Analysis)

Probably the most important criterion of a front in flat terrain (oceans included) is the cyclonic wind shift. Do not be misled by nonrepresentative winds such as along the coast, large lakes, in valleys, etc., and directions of light winds (under 10 knots) on morning maps.

For a moving front, 3-hour pressure tendency differences over land are very helpful, especially in mountain areas. From a moving ship, they are of no value unless a dependable correction for the movement of the ship is made.

Pressure tendency characteristics may be of little help, particularly in summer. Characteristics of the barograph trace (3-hour) may be helpful when the magnitudes of the changes are large by comparison with diurnal effects, dynamic effects of convective activity, etc. The test of usefulness is if the particular characteristic is organized along a line.

Actual values of temperature and dewpoint are likely to be unrepresentative near a front. Temperature or dewpoint gradients are more reliable with weak gradients (homogeneity) in the warm air mass and strong gradients on the cold air side of the front. Due to horizontal mixing, heating or cooling from below, and evaporation from below near the shallow edge of the cold air mass, temperatures and dewpoints on the edge close to the cold air side of the front may closely approach those of the warm side. In these circumstances the station's wind direction, if representative, is usually the determining factor. Surface temperatures are generally of little value in frontal analysis on morning maps, and are most representative on afternoon or evening maps.

AIR MASS ANALYSIS

While an Aerographer's Mate familiar with a given weather situation may have little need for air mass classification, the distinction between air masses of different types is nevertheless important to weather analyses and is a means of conveying a more complete picture to a user of the weather map, such as a pilot.

If masses of air of different characteristics are isolated from one another, the transition zones between them can be determined as the first step in frontal analysis. The reverse procedure may be used to advantage in examination of vertical temperature soundings, in which case the air masses over a station may be isolated from one another by locating the transition zones or frontal discontinuities on the soundings.

On the basis of hydrometeors, such as fog or drizzle and cloud types, a stable air mass can often be differentiated from an unstable air mass across a narrow zone. In considering the characteristics of air masses according to accepted classification, many modifications have to be taken into account, particularly in the warm air mass where, for example, stratus may form with rapid movement over a cold surface while fog will tend to form if the flow is sluggish. In warm seasons, the warm air mass can change from foggy at night to unstable in the daytime. At sea the diurnal influence is slight, with a tendency for greater stability in the daytime in lower layers.

Characteristics of the same general type of air mass vary to some extent with their source region, Arctic air, for example varying somewhat, depending on whether it has developed in Siberia, over the polar cap, or in Canada.

No definite values of surface temperature or dewpoint can be stated for continental polar air masses, since these depend upon source region, length of time over the source, depth, trajectory, heating and cooling from below, and the like. Often there are marked horizontal gradients of temperature within the air mass. The amount by which a cold air mass is warmed, both surface aloft when passing over a warmer surface, is greater when there is subsidence aloft.

Arctic air at midlatitudes generally has a surface temperature of about 0°F or below. The only air mass which in midlatitudes has approximate homogeneity at the surface and to some extent aloft is maritime tropical air. In summer, mT air over the contiguous United States has a representative (maximum) surface temperature of near 90°F and a dewpoint of around 70°F. Corresponding values in winter are roughly 75°F and 60°F.

PRECIPITATION ANALYSIS

Ordinarily, precipitation analysis on surface charts is not completed before transmission of the coded maps (at weather offices that transmit canned maps), because of insufficient time. It is, however, carried out later and serves as an aid to the subsequent analysis for which precipitation reports may then be more quickly interpreted.

Steady precipitation is shaded in green. Shower areas are marked with shower symbols; thunderstorm area, with thunderstorm symbols; and drizzle area, with drizzle symbols, in accordance with standard practices. The edges of an altostratus shield and fog areas may be outlined.

Warm-front rain, or a corresponding steady rain area in advance of an occluded front, is usually assumed in the absence of data to extend over a band of approximately 300 to 350 miles ahead of the front. This typical model may of course be modified by geographical factors or complicated frontal structure. A known boundary of solid altostratus will often indicate that the edge of the precipitation area is near. Also, the presence of "mid 7" clouds frequently shows that a rain area is approaching. The existence of this cloud type, or of an altostratus layer, may be the first indication of a front approaching a coastline.

Pressure falls in an area of precipitation may indicate that the precipitation is a warm-front type, the fall of pressure tending to be greatest where the rainfall is heaviest.

If the band of precipitation is appreciably wider than 300 to 350 miles, a double front structure may exist, or there may be a smaller than normal slope to the warm-front surface. Sometimes there is an increase of slope well in advance of the surface position, in which case the position of the line along which the front begins to increase in slope may, if considered significant, be shown as a warm front aloft. Where relatively dry air ascends over a warm front, precipitation is often lacking, or may occur only some distance ahead of the surface position of the front.

Shower symbols are used to give form to the region where present weather, weather in the last hour, past weather, precipitation amounts, or the presence of cumulonimbus suggests showery conditions, or where, in the absence of data, shower conditions are expected on the basis of continuity or model.

In cold air masses, showers will be prevalent where there is rapid warming over water surfaces as over the Great Lakes in winter; over mountainous country, such as the western slope of the Appalachians, and in general where the isobars are curved cyclonically and the cold air contains sufficient moisture.

Drizzle is looked for principally where warm, moist air moves rapidly over a cooler surface, especially in the warm sector of a cyclone. It may occur in cold air, especially under anticyclonic or stagnant conditions if there is sufficient moisture and instability in the lower levels and a stable layer or inversion above the low clouds.

The principal cloud or precipitation systems looked for in a typical extratropical cyclone are:

1. Warm-front rain areas, and the prewarm-front cirrus and altostratus preceding them.
2. Warm sector drizzle, low clouds, or fog.
3. Warm sector showers: in many cyclones most showers occur in the warm sector rather than along the cold front.
4. Cold-front (squall line) cumulonimbus and showers.
5. Post-cold-front cumulus and showers (usually only where flow is cyclonic or showers are favored by topographic features).

All systems are subject to modifications by geographical features, moisture content of the air, and other characteristics pertaining to the given cyclone.

ISALLOBARIC ANALYSIS

The isallobaric field can be useful in early detection of frontogenesis, and also in determining the direction of movement of a front where weak wave activity is taking place, where the actual wind at gradient level is different from that indicated by the isobars, or where questionable reduction factors make sea level isobars unrepresentative. In addition, the instantaneous movement of high and low centers is apparent when isallobars are drawn. The isallobaric field is frequently of value in indicating various types of developments, such as the formation of squalls or genesis of the instability line in the warm sector of a cyclone and, more important, changes in the structure or movement of pressure systems. Three hourly tendencies are more significant if the diurnal correction is applied.

FORMATION AND DISSIPATION OF FRONTS

The map should be scanned for indications of frontogenesis and frontolysis in accordance with the information contained in chapter 5 of this training manual. Keep in mind that certain geographical areas during specified periods of the year are favorable for the formation or dissipation of fronts. Also, certain portions of the frontal system itself are favorable for frontogenesis, such as the trailing edges of long cold fronts whose orientation has become east-west and the isobars either anticyclonically curved or parallel to the front.

Indicate the areas with the appropriate symbols: "FG" for frontogenesis and "FL" for frontolysis.

APPLICATION OF SATELLITE PHOTOGRAPHS

As mentioned previously in this chapter, satellite photographs are a valuable aid to the analyst. The photographs should have been referred to when initially positioning the fronts, lows, troughs, and ridges on the map. Prior to finalizing the map these pictures should be compared with the analysis to insure that the complete analysis is a consistent, smooth-flowing picture. It may be desirable to make last minute adjustments inadvertently overlooked during the initial phase of the analysis.

APPLICATION OF COMPUTER PRODUCTS

The application of computer products was presented earlier along with references for obtaining the various charts. Just as it is important to compare the satellite pictures, the computer products used during the initial phase of the analyses should also be compared with the finished map prior to finalizing. As stated earlier, in many instances these charts will be utilized in place of the hand drawn analysis due to personnel limitations and limitations in the availability of data.

FINALIZING THE CHART

Some of the steps outlined below will already be completed or in the final stages by the time you have reached this step in the analysis. You should either complete or check for completion the following elements of the analysis:

ISOBAR LABELING

After all the isobars have been sketched and the position of the fronts located, the final analysis is made. Isobars should be erased and redrawn, smoothing out irregularities which do not show logical consistency and other irregularities which may be due to errors in reported pressure values.

FRONTS

After completing the isobaric analysis, color in the fronts, using the appropriate standard symbols and indicating areas of frontogenesis and frontolysis.

HIGHS AND LOWS

Label the centers of highs with a blue "H" and the centers of lows with a red "L." If the center of a high is weakening (central pressure falling) or the center of a low is deepening (central pressure falling), suffix the letter "L" or "H" with a minus sign. If the central pressure is rising, use a plus sign.

WEATHER PHENOMENA

Precipitation areas and other areas of significant weather should be indicated by using the appropriate color shading or symbols.

AIR MASS LABELING

This is usually an optional entry and many weather offices no longer enter the types of air masses on their base charts. If this step is desired, label the air masses in accordance with the standard air mass symbols. Sometimes an alternate method is used to indicate the air masses. This method either shades the ~~cold~~ and warm air masses in light blue and red, respectively, or is indicated by a large arrow, shaded in lightly and labeled appropriately, showing the direction of movement of the air mass.

FUTURE MOVEMENT OF FRONTS AND PRESSURE SYSTEMS

Expected positions of fronts and pressure systems may be indicated by the use of arrows pointing to the position they are expected to move during the next 6, 12, or 24 hours.

ISALLOBARS

As pointed out previously, isallobars are most representative if the diurnal characteristic is removed. Isallobars are usually drawn for 1 millibar intervals, 1, 2, 3, etc., plus and minus tendency values. Dashed blue lines are used for plus or positive tendencies and dashed red lines for minus or negative tendencies. Some stations prefer to begin with either the plus 2 or minus 2 value, thereby eliminating some of the diurnal change. The fallacy of this is that while this may be representative during certain periods of the day when either plus or minus diurnal changes correspond with these changes in the synoptic tendency, you may be removing a perfectly valid tendency change. It is best to use diurnal tendency tables or charts as mentioned in a previous section of this chapter.

SOUTHERN HEMISPHERE ANALYSIS

Most of the Aerographer's Mates have a Northern Hemisphere orientation and all too often completely ignore the peculiar aspects of the Southern Hemisphere until they are called upon to make analyses in this area. It requires a considerable amount of effort to reorient yourself to the fact that cyclonic circulations in this hemisphere have a clockwise rotation of the winds and anticyclonic circulations are counterclockwise. Also, the pressure patterns are more regular in this hemisphere due to the absence of the large land bodies, present in the Northern Hemisphere. However, certain similarities do exist between both hemispheres in that all atmospheric features which depend fundamentally on static relationships, the hydrostatic equation in particular, are the same. For example, in both hemispheres fronts, troughs and lows slope toward the coldest air, ridges and highs toward the warmest air, isobar kinks point toward higher pressure, and static stability criteria in the form of lapse rate and moisture relations are unchanged. The basic differences of hemispheric analysis can be subdivided into two groups: geographical (including topographical), and dynamical.

GEOGRAPHIC CONTRASTS

Aside from the greater land area of the Northern Hemisphere, the distribution of continental areas (including ice surfaces which do not break at any season) is quite different. Only within 10°-25° of the Equator are the landmass areas of the two hemispheres comparable. The great landmasses of the Northern Hemisphere extend from subtropical to subarctic latitudes, with the Arctic Ocean covering most of the area north of the 65th parallel. In the Southern Hemisphere only one-fifth of the 30th parallel crosses land and only one twenty-fifth of the 40th is not maritime. From 45° S to 65° S there is virtually no unbroken water, with the edge of the Antarctic icecap oscillating seasonally near the latter circle.

In polar regions, the Arctic icecap seldom rises more than a few feet above sea level, and areas of open water appear during the summer season. The topography of the Antarctic icecap rises to altitudes over 13,000 feet. Most of the area is a plateau of mean elevation of about 10,000 feet, dropping sharply to sea level around the periphery. Only a few isolated areas of open water have ever been observed in summer. Greenland is the nearest North American analog.

The longitudinal distribution of land and water is also different, with two major continents and oceans in the Northern Hemisphere and three each in the Southern Hemisphere. The principal topographical features of 50° S latitude are the Andes of South America and the plateau of South Africa, with its mean elevation about 5,000 feet, dropping rather abruptly to sea level all around the periphery as does the Antarctic icecap. Australia is practically without marked topographical features, and nowhere south of the Equator are there any features comparable to the Alps, Urals, or Himalayas.

North, as we think of it in the Northern Hemisphere, has a different connotation in the Southern Hemisphere. North means equatorward in this hemisphere and south means poleward.

DYNAMICAL CONTRASTS

All dynamical differences stem from one fact; the coriolis parameter is negative in the Southern Hemisphere. For example, the basic relationship between wind and pressure as expressed in Buy Ballot's law becomes in the Southern Hemisphere: Facing in the direction toward which the wind is blowing, low pressure is found on the right. Consequently, the sense of rotation about lows is clockwise and about highs is counterclockwise in this hemisphere. However, lows are called cyclones and highs anticyclones in either hemisphere.

Cyclonic vorticity is negative and anticyclonic vorticity is positive in the Southern Hemsiphere. The sign conventions for divergence and convergence are unchanged, but relations between divergence and vorticity must be modified. All statements relating temperature and changes in wind with height must be altered for the Southern Hemisphere since they involve applications of Buy Ballot's law to thermal winds and temperature fields.

GENERAL CIRCULATION

Hemispheric differences, in general, decrease with increasing altitude. All the jet streams of the Northern Hemisphere have their southern analogues. Southern jets are more intense on the average with smaller amplitudes, reflecting the greater zonal indices (over double in magnitude) of this hemisphere. The stratospheric cyclone of the polar night transforms into the anticyclone of the polar day in both hemispheres, but the "explosive warming" prior to the onset of the polar day has not yet been observed over the Antarctic.

Blocks are comparatively rare and occur southeast of continents in late winter and early spring, and drift slowly eastward, again emphasizing the importance of middle latitude continents on features of the general circulation.

There is some doubt as to whether counterparts of the Aleutian and Icelandic lows exist in the mean circulation of the Southern

Hemisphere. If they do, they are located in the Ross and Weddell Sea areas, on the edge of Antarctica, and are most sharply defined at about 700 millibars. Navy meteorologists with experience in those areas indicate that many lows undergoing cyclolysis move into these regions but disappear rapidly at low levels, retaining their identity as cold lows aloft for longer periods. Tentatively, it can be said that the Ross and Weddell Sea lows exist aloft in both the general and mean circulation, but near sea level only in the general circulation. Not all lows are stopped on the periphery of Antarctica; they penetrate to the South Pole.

The subtropical highs of the South Atlantic, South Pacific, and South Indian Oceans differ from their northern counterparts in number, permanence, seasonal migration, and seasonal intensity changes. They are more migratory zonally, less migratory meridionally, and they decrease in intensity from winter to summer.

MEAN PRESSURE CHARACTERISTICS

The mean pressure distribution over the Southern Hemisphere shows three large semipermanent highs exist in the midlatitudes over the oceans. (See fig. 4-5.) One is in the eastern South Pacific extending from about 140 degrees west to the west coast of South America; a second one almost completely covering the South Atlantic Ocean; and the third more or less centrally placed over the South Indian Ocean. Along the mean axes of these highs, about 30°S and between these highs, lie indifferent regions of cols. South of this high pressure belt, pressure decrease is regular and marked to about 65° south where a continuous low pressure trough known as the "Antarctic Trough" encircles the Antarctic continent. Farther south is found higher pressure, over Antarctica. North of the high-pressure belt, from 30° south, is the region of low pressure, located over the equatorial regions, that separates the high pressure belts of both hemispheres. In summer heat lows appear south of the Amazon Basin in South America, over the whole of the eastern part of South Africa, and over northern Australia. The latter is the largest of these heat lows and covers northern Australia, the Dutch East Indies, and the central Indian Ocean. In winter, although pressure is relatively high over land, separate subtropical anticyclones can still be distinguished. The only pronounced seasonal variation of pressure in midlatitudes is the shift of the axes of the subtropical highs, moving only a few degrees of latitude northward during the Northern Hemisphere summer and a few degrees of latitude southward in the Northern Hemisphere winter.

AIR MASSES

The semipermanent highs produce tropical air similar to the highs in the Northern Hemisphere. There is no continental polar air in the Southern Hemisphere. The air over the snow and ice covered regions is Antarctic air but it rarely leaves this area as true Antarctic air. It becomes rapidly modified to mP air as it moves over the water. Maritime polar air is the most predominant air of the Southern Hemisphere. Air forming in the central dry regions of South Africa and Australia has all the properties of continental tropical air and, in its source region, is dry and cloudless with a steep lapse rate. When it moves to neighboring oceans, a strong inversion is formed in the lower layers and with the addition of moisture from the ocean, sheets of stratocumulus and stratus are formed.

MAJOR FRONTAL ZONES

In figures 2-7 and 2-8 you can see the major frontal zones of the Southern Hemisphere. Areas of frontogenesis are found in the semistationary polar troughs that exist along the western border of each of the subtropical highs. Wave disturbances form along these fronts much the same as those forming along the polar fronts in the Northern Hemisphere. Usually they develop into families of from two to six. This wave sequence ends when the final wave member of the series has run together and occluded in a large central cyclone to the southeast of the semipermanent highs.

SYNOPTIC CHARACTERISTICS OF THE PRESSURE PATTERN

Between the regions of semipermanent highs are trains of warm migratory anticyclones. They move steadily at about 10 degrees of longitude per day from west to east. Between these migratory highs are the upper parts of inverted V-shaped depressions that move along with them. These V-shaped depressions are the northern parts of larger cyclonic systems to the south.

APPLICATION OF SATELLITE CLOUD PHOTOGRAPHS

The application of satellite cloud photographs to analysis and forecasting in the Southern Hemisphere utilizes the same procedures as presented earlier in this chapter for the Northern Hemisphere. The analyst must bear in mind, however, that the orientation of the cloud systems in relation to the meteorological features must be adjusted in accordance with the reversed flow.

APPLICATION OF COMPUTER PRODUCTS

As mentioned earlier in this chapter, computer products will frequently be used in place of the hand drawn analysis. This practice is rapidly becoming standard procedure, and is even more applicable to the Southern Hemisphere due to scarcity of data. It is therefore imperative that the AG learn as much as he can

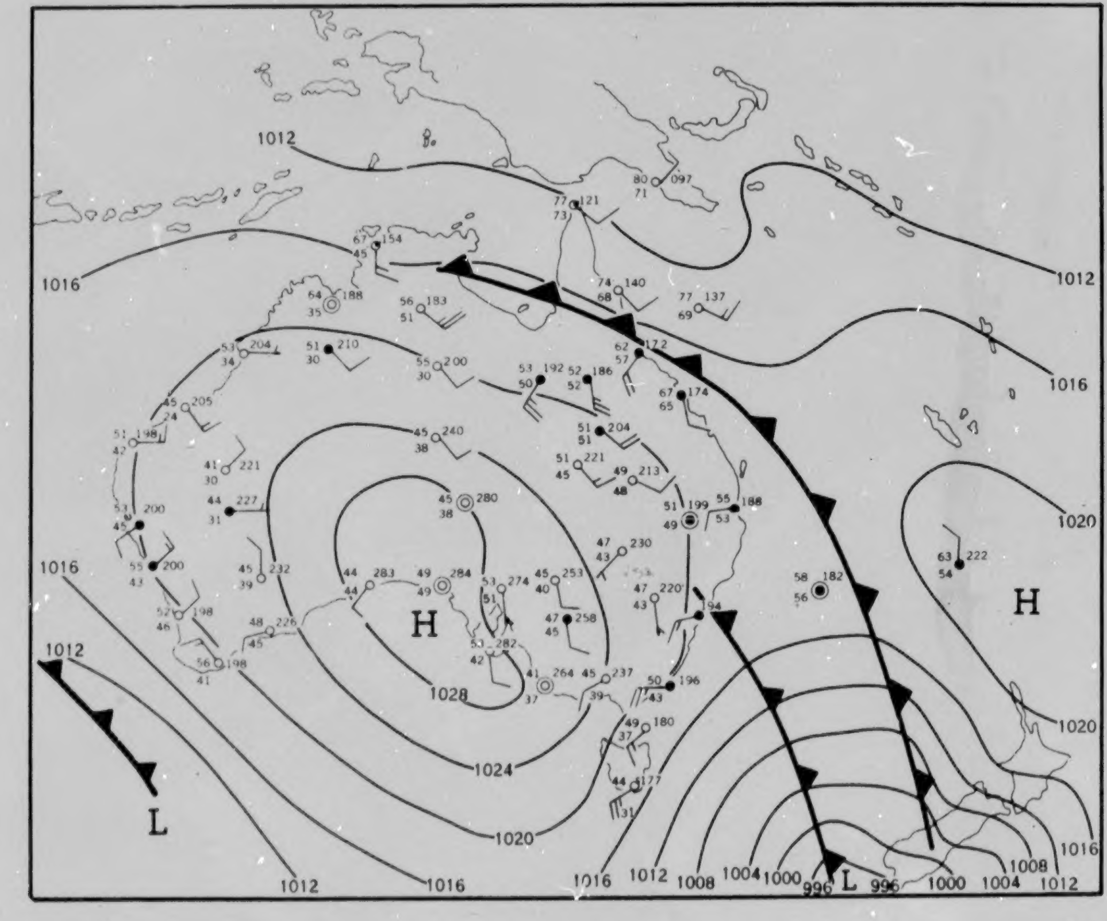

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Figure 6-19.—Weather map for a portion of the Southern Hemisphere.

about the application of computer products in analysis and forecasting. Their application will be discussed in greater detail in chapters 8 through 12 pertaining to forecasting. The references mentioned earlier in this chapter as well as any later or more current publications should be carefully examined for additional information.

SYNOPTIC ANALYSIS

Basically, the techniques of Southern Hemisphere analysis do not differ appreciably from those of its Northern Hemisphere counterpart, but because the Southern Hemisphere is largely made up of water, oceanic analysis is of greater importance. There are some notable exceptions. Since warm type occlusions develop principally off the west coasts of continents in middle latitudes, it is extremely unlikely that such occlusions ever occur in the Southern Hemisphere. The so called "meridional fronts" are usually cold occluded fronts or sometimes just cold fronts. Fronts are usually weaker in density contrast than in the Northern Hemisphere. Figure 6-19 shows a typical analysis of surface chart in the region of Australia. Note that the feathers are drawn on the right side of the shaft looking toward the station circle.

Some basic aids to analysis over the Southern Hemisphere are listed below:

1. Extensive use of wind scales should be utilized over the water areas.
2. A knowledge of terrain and geographic characteristics is important when considering island and other land reports.
3. Seasonal variations that may indicate erroneous heat lows must be handled with caution.
4. Because of the rapid altitude changes in short distances on the edges of South Africa and Antarctica, no attempt should be made to connect the pressure fields of the continental and adjacent ocean areas. Discontinuous isobars are in order wherever terrain heights differ by 5,000 feet or more.

CHAPTER 7

UPPER AIR ANALYSIS

Practically all weather as we know and experience it occurs in the troposphere. For this reason the Aerographer's Mate must have data from levels other than the surface to provide himself and other users with the most complete picture possible of the 3-dimensional distribution of wind and temperature in the atmosphere. Upper air observations yield most of these data and when plotted and analyzed on upper air charts and diagrams, these analyses can be related to the surface and other upper air manifestations to gain a more accurate picture of the existing conditions.

EVALUATION OF DATA

DATA CONSIDERATIONS

The analyst should be aware of, and alert for, discrepancies between reported data and that actually existing at the particular level. The same type of errors and unrepresentativeness of data can exist at upper levels as at surface levels and depend to a large extent on the method used in determining the data.

Temperature

An international comparison of radiosondes sponsored by the World Meteorological Organization in Switzerland in 1950 revealed that in 95 percent of the cases the differences between the data given by any two radiosondes on the same flight were not greater than 2°C and 20 percent relative humidity up to 700 mb. The results showed that while some instruments are affected too greatly by solar radiation and more accuracy is needed in humidity measurements below 0°C, the instruments provide satisfactory data for synoptic use if observers are fully trained to exercise the proper care in the evaluation of soundings. The compatibility tests showed that in practice an individual radiosonde will report temperature to within 1°C, 91 percent of the time.

Pressure

The average instrumental error in pressure is from 2 to 4 mb with an absolute maximum of about 5 mb. This maximum error, however, is found in only 1 out of 100 instruments. The mean virtual temperature estimate of an isobaric layer of any appreciable thickness is, for all practical purposes, unaffected by the relatively small instrumental error.

Humidity

In the average upper air sounding the humidity reports are about 5 percent in error at the 700-mb level, and they may be as high as 20 percent at the 300-mb level. Relative humidity values are always given with respect to a water surface and never with respect to ice crystals. This means that in passing through a cirrus type cloud, the humidity element may give only a 70 to 80 percent relative humidity and still be accurate. For example, at −20°C the humidity with respect to water need only be about 83 percent for the air to be saturated with respect to ice. In this respect one can make a very rough estimate that for every 10°C below zero, a 10 percent lowering of the reported relative humidity can be anticipated due to the fact that the humidity is coded with respect to water rather than with respect to ice.

Upper Winds

The reliability of upper wind observations depends upon the manner in which they were determined. For example, when plane triangulation is used to determine the height of the balloon, and the curvature of the earth is not taken into account, the balloon seems to be lower than it actually is.

Wind directions are often very unreliable when light winds in the stratosphere are observed above strong winds in the troposphere, and they may be as much as 180 degrees off in direction.

Summary

Table 7-1 is the table used by the National Meteorological Center for the probable error in height, wind, and temperature data at various elevations.

ORDER OF ACCURACY OF UPPER AIR DATA

The order of accuracy of upper air data is listed below.

Rawinsondes (Radiosondes)

Rawin wind data are for the most part very accurate. Heights and temperature data are subject to the same type errors listed previously. Wind data are most accurate when determined by radar.

Pibal Data

The outstanding accuracy limitations of this type of observations are the assumption of a constant ascension rate and the possibility of errors because of the timelag in readings by the manual operator. Errors in pilot balloon data may become especially significant in the case of strong winds or large vertical shears.

Weather Reconnaissance Reports

Pressure, temperature, and humidity observations gathered by dropsonde methods are limited somewhat in accuracy by the types of instruments used. In general they are not as accurate as rawinsonde and radiosonde equipment. However, flight altitude data are reasonable accurate and useful. Heights are usually given at some mandatory pressure surface (700-, 500-mb, etc.).

AIRCRAFT REPORTS (AIREP'S)

The accuracy of AIREP data will vary radically with the method of position determination, the instruments available to the navigator, instrument limitations, and the navigation procedures used. When extrapolations are made from flight altitude to some other level, these must be corrected. In those cases where the mean wind over the last hour is reported, the wind should be spotted at the midpoint of the last hourly track.

RESOLVING ERRONEOUS DATA

Reports are occasionally received and plotted which are evidently not compatible with other observations in the same region. In many cases such erroneous reports can be resolved into usable data. Such errors may appear for a number of reasons. Common ones, which can sometimes be resolved, are communications and plotting errors, computation errors, and in the case of aircraft reports, erroneous position reports. Although the Aerographer's Mate must exercise discretion in correcting such errors, he should not throw away or disregard data, unless it is absolutely necessary.

PURPOSE OF ANALYSIS

The purpose of the analysis of upper air charts is to estimate and to represent the continuous distribution of atmospheric conditions throughout three dimensions from observations made at but few points or along a few lines. Such an analysis should provide pictorial representations of the conditions everywhere; otherwise the many interrelationships between the various conditions to be considered would be too difficult to be useful.

In many regions the greatest amount of observational data is available at the surface of

Table 7-1.—Probable errors in upper air data.

Parameter	Probable error	Remarks
Raob temperatures	± 1° C to 400 mb ± 2° C above	
Raob heights 1,000 mb	± 6 meters per 305 meters of station elevation	Different temperature parameter used for reduction to 1,000 mb than to msl which results in inconsistencies of the order of 120 meters in extreme cases. In order to circumvent this, we directly convert the reported msl pressure to 1,000 mb.
850 mb	± 9 meters	
700 mb	± 12 meters	
500 mb	± 21 meters	
400 mb	± 30 meters	
300 mb	± 49 meters	
200 mb	± 70 meters	
150 mb	± 90 meters	
Thicknesses:		
1,000-700	± 12 meters	
1,000-500	± 21 meters	
500-300	± 27 meters	
Tropopause heights	± 10 mb	
Winds aloft directions	± 5 degrees	Due in part to rounding off to nearest 5 degrees in coding.
Winds aloft speeds	± 10 knots up to 500 mb for winds up to 50-75 kt	This error increases rapidly, especially for winds over 100 kt, usually due to low elevation angles of recording equipment. Also varies considerably with different types of equipment.

the earth. In those cases it is desirable to start with the surface level and work up through the successive upper levels by the use of thickness analyses. This method is known as differential analysis and is discussed later in this chapter. However, in some regions data are adequate and ample at the upper levels and only some intermediate points must be constructed.

If the complete 3-dimensional picture is available to the meteorologist, his forecast problem is greatly simplified. Therefore, it is important that data be evaluated, extrapolated, and analyzed accurately to portray the best possible picture of the actual atmospheric conditions. Some of the data useful to both the meteorologist and the pilot are apparent from a preliminary study of the map. Other data must be computed or derived from this. Two of the most important considerations are the use of winds, both actual and computed, in the analysis of contours, and the extrapolation of heights in areas where data are sparse.

Constant pressure charts are used for:

1. Locating pressure systems.
2. Steering pressure systems.
3. Locating dry and moist layers of air.
4. Determining if fronts extend to the level in question.
5. Locating cyclonic and anticyclonic flow.
6. Locating areas of horizontal convergence and divergence.
7. Forecasting surface and upper air weather.
8. Constructing thickness charts and advection charts.
9. Constructing time differential charts.
10. Jetstream and isotach analysis.

EXTRAPOLATION OF HEIGHTS OF CONSTANT PRESSURE SURFACES

In practice, the analysis of upper level charts may be conveniently placed in two categories—the analysis of sparse data areas and the analysis of dense data areas. On many occasions, upper air observations do not reach all the way up to the level where an analysis must be made. Because of the scarcity of upper air reports over ocean areas, it is frequently desirable to extrapolate upper air data from surface reports. There are many methods, employing the use of tables or nomograms, which may be used to accomplish this purpose. All of these are based on some form of the hydrostatic equation. With the known surface temperature and pressure, and the assumed mean virtual temperature to the desired level, it is possible to obtain the height of that level. The mean virtual temperature is usually obtained by assuming a moist adiabatic lapse rate or by averaging the known surface temperature and an estimated temperature for the upper level. The latter estimated temperature is based on previous analysis compensated for any changes which may have occured during the interim. In the case of marked inversions of any type, the estimated height is less than the "true" height since the arithmetic mean of the temperatures will then be less than the "true" mean temperature. Of course the analyst can compensate for such inversions when their presence is suggested on the synoptic map by using a high estimate of the upper temperatures. This procedure could be used in the following situations.

1. In the vicinity of high pressure cells where subsidence inversions are present.
2. When surface inversions are indicated by stable weather phenomena such as fog.
3. When a frontal surface is below the level to be extrapolated.

A recent study comparing the true thickness of the 1,000-700-mb and 1,000-500-mb layers to the value obtained by using the arithmetic mean of the 1,000-mb temperature and the 700- or 500-mb temperature, respectively, showed an error of less than 30 meters in 90 percent of the cases for the former, and less than 60 meters in 90 percent of the cases for the latter. This sample included several hundred soundings.

Obviously, if we are to construct the height of the 700- or 500-mb level, the first correction that must be made is that of correcting the current temperature and pressure to the 1,000 mb level and then extrapolating the height of the upper level.

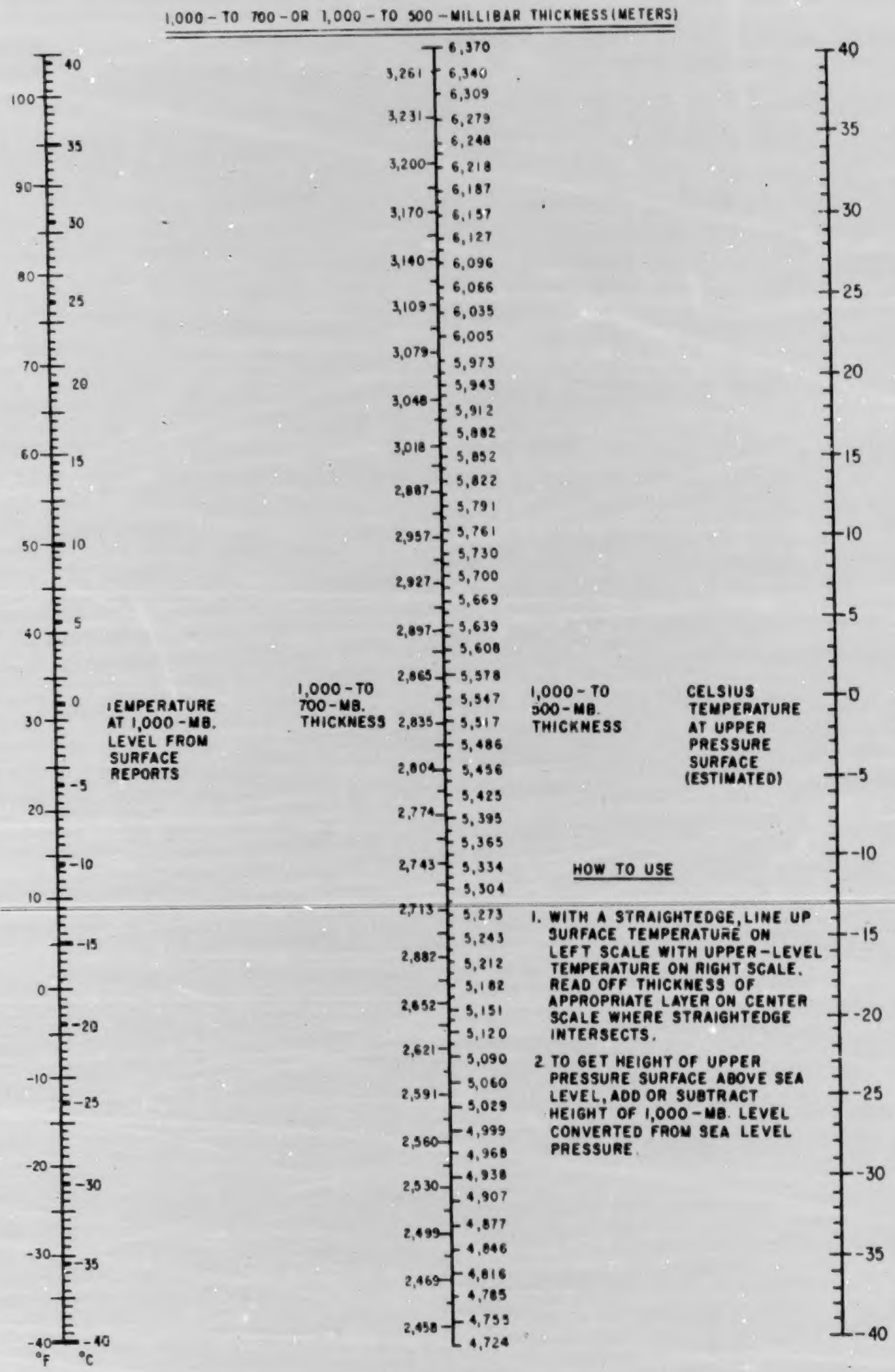
Computing Height of the 1,000-mb Surface

The height of the 1,000-mb pressure surface is the height of this surface in geopotential meters above mean sea level. For computational purposes, assume that 7 1/2 mb equals 60 meters (or 8 meters per 1 millibar) when determining this height.

Extrapolation of Upper Levels

One of the methods for obtaining the height of the 700- or 500-mb level is outlined below:

1. Estimate the temperature at the upper level from past analysis.
2. Use figure 7-1. With a straightedge line up the surface temperature on the left scale with the upper level temperature on the right scale.



HOW TO USE

1. WITH A STRAIGHTEDGE, LINE UP SURFACE TEMPERATURE ON LEFT SCALE WITH UPPER-LEVEL TEMPERATURE ON RIGHT SCALE. READ OFF THICKNESS OF APPROPRIATE LAYER ON CENTER SCALE WHERE STRAIGHTEDGE INTERSECTS.
2. TO GET HEIGHT OF UPPER PRESSURE SURFACE ABOVE SEA LEVEL, ADD OR SUBTRACT HEIGHT OF 1,000-MB. LEVEL CONVERTED FROM SEA LEVEL PRESSURE.

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Figure 7-1.—Nomogram for computing height of the 700-mb and 500-mb levels.

Read off the thickness of the appropriate layer where the straightedge intersects the center scale. Note that this is the thickness from 1,000 mb to the upper level and that values for the 1,000- to 700-mb thickness are on the left of the center scale and values for the 1,000- to 500-mb thickness are on the right of the center scale.

3. Estimate the height of the 1,000-mb level.

4. Add algebraically the height obtained in step 3 and the thickness obtained in step 2. This is the height of the upper surface. For example, a ship reports a pressure of 1,015 mb and a temperature of 78° F. If the temperature of the 700-mb level is estimated to be -5° C, what is the height of the 700-mb level?

a. The 1,000- to 700-mb thickness equals 2,957 meters (from fig. 7-1).

b. The height of the 1,000-mb level equals 120 meters (1,015 mb minus 1,000 mb equals 15 mb; 15 mb divided by 7 1/2 mb equals 2; 2 times 60 meters equals 120 meters if the standard value is used).

c. The height of the 700-mb level equals 2,957 meters plus 120 meters equals 3,077 meters.

In areas of sparse reports and if a quick computation of the thickness is desired, figure 7-2 may be used to compute the thicknesses between the 1,000-700 mb layer and the 1,000-500 mb layer using the temperatures at the 850 and 700 mb levels, respectively. A column is also included in this table to show the thickness from 500 to 300 millibars based on the temperature value at 500 millibars. A word of caution in using this table: inversions or nonrepresentative temperatures at either of these levels will result in an incorrect thickness for the layer. For a more correct computation based on this method see Table 53, Smithsonian Meteorological Tables, NA 50-1B-521.

	Z		Z		Z
t850	1000-700	t700	1000-500	t500	500-300
-38	2439	-41	4694	-47	3261
-35	2469	-37	4755	-46	3292
-32	2499	-34	4816	-45	3292
-27	2560	-31	4877	-43	3322
-24	2591	-28	4938	-41	3322
-21	2621	-25	4999	-39	3353
-16	2682	-23	5060	-37	3353
-13	2713	-20	5121	-35	3413
-10	2743	-17	5182	-33	3444
-7	2774	-14	5243	-30	3475
-4	2804	-11	5304	-27	3505
-1	2835	-9	5365	-24	3536
+1	2865	-6	5425	-22	3566
+4	2896	-3	5486	-20	3597
+7	2926	0	5547	-19	3627
+12	2987	+3	5608	-17	3658
+15	3018	+5	5669	-14	3688
+18	3048	+8	5730	-11	3719
+22	3079	+11	5791	-9	3749
+26	3139	+14	5852	-7	3780
+30	3170	+17	5913	-6	3810

LEGEND ON TABLE.

Z's - thickness in meters between specified constant pressure levels

t's - temperatures in degrees Celsius at specified levels

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Figure 7-2.—Thickness values from temperatures at constant pressure levels.

EVALUATION OF WIND

When evaluating data during the analysis of upper winds, the analyst should differentiate between that data which is related to actual wind flow as compared to the flow of geostrophic wind.

Differences Between Actual and Geostrophic Winds

It is sometimes helpful to consider some theoretical ways in which the geostrophic wind should differ from actual winds. At low levels (below 2,000 feet above the terrain) friction with the ground is effective and actual winds should flow across contours or isobars from high to low values. Above the friction layer, discrepancies should occur whenever the flow is curved rather than along straight lines. Actual winds should be somewhat weaker than geostrophic winds in counterclockwise flows around lows, and somewhat stronger than geostrophic winds in clockwise flows around highs. In low latitudes (between 20° N and 20° S) geostrophic winds become poor estimates of the actual winds because the geostrophic wind is inversely proportional to the sine of the latitude which approaches zero near the Equator.

Wind Shear

Wind shear may be either on a vertical or horizontal plane. The vectoral rate of change of wind with respect to altitude is called the vertical wind shear. Horizontal wind shear is the rate of change on a horizontal plane. The vertical wind shear per unit distance within a layer of air may be determined by taking the vector difference between the wind reported at the top of the layer and the wind reported at the bottom of the layer and dividing by their vertical separation.

Thermal Wind and Wind Shear

The vector difference between the geostrophic winds at two levels is called the thermal wind by meteorologists. Consequently, the thermal wind is not a wind which actually blows in the atmosphere; rather, it is the difference between winds at two levels. To illustrate, if the 700-mb geostrophic wind is southwest at 50 knots, and the 500-mb geostrophic wind is west at 80 knots, then the vector difference or the thermal wind is northwest at 58 knots. (See fig. 7-3.)

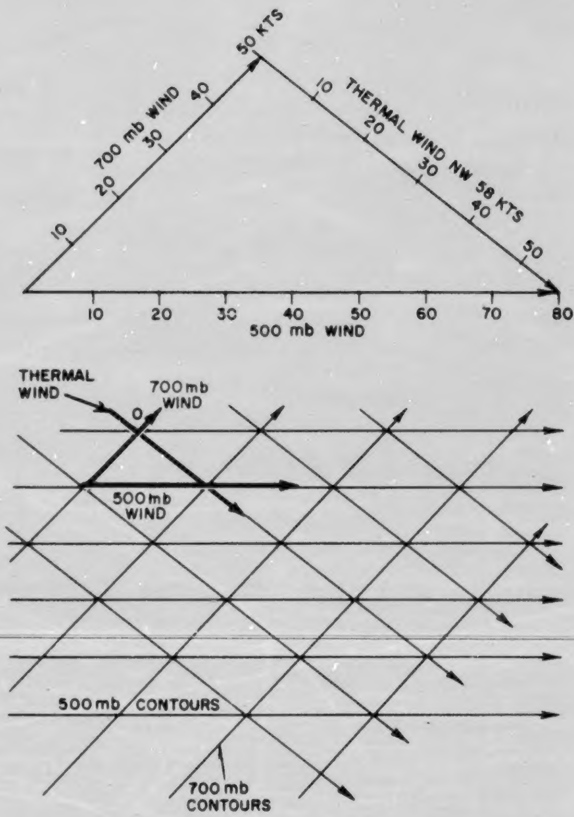

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Figure 7-3.—Thermal wind between two pressure surfaces.

In this example, the mean isotherms between 500 and 700 mb would be oriented in a northwest to southeast direction with colder air toward the north. In general, the direction of the thermal wind is parallel to the mean isotherms in the given layer, with the colder air to the left; looking downstream. The magnitude of the thermal wind is directly proportional to the mean temperature gradient of the layer.

In summary, the thermal wind bears the same relationship to the horizontal mean temperature gradient in a layer of air as the geostrophic wind does to the horizontal pressure gradient.

One must distinguish between the change of the actual wind through a layer and the thermal wind of that layer. The actual wind change is the vector difference of the observed winds at the two levels. On the other hand, the thermal wind is a special type of wind change vector; if the winds at both levels are geostrophic, the actual wind change and thermal wind are identical. In some cases, as when the curvature of the contours changes rapidly with height, there may be a difference between the thermal wind and the actual wind shift through the layer.

One of the most fundamental meteorological concepts used by meteorologists in upper level analysis is the relationship between wind and height as expressed by the geostrophic and gradient wind equations.

Geostrophic Wind Scales

The mathematical equations representing deviations in geostrophic wind are applied to analysis through the use of scales. The procedure for using these scales is presented in chapter 6 of this manual. The use of geostrophic wind scales makes practical application of the equations possible within the time-frame necessary to maintain validity in the analysis.

APPLICATION OF SATELLITE DATA

The cloud formations depicted on satellite pictures have a definite relationship to the flow patterns which appear through the analysis of computer or hand drawn weather charts. Figure 7-4 illustrates the relationship between an upper air analysis and an HRIR (high resolution infrared) satellite picture showing a cutoff low at the 500-mb level. Figure 7-5 shows the same system appearing in an AVCS satellite photograph.

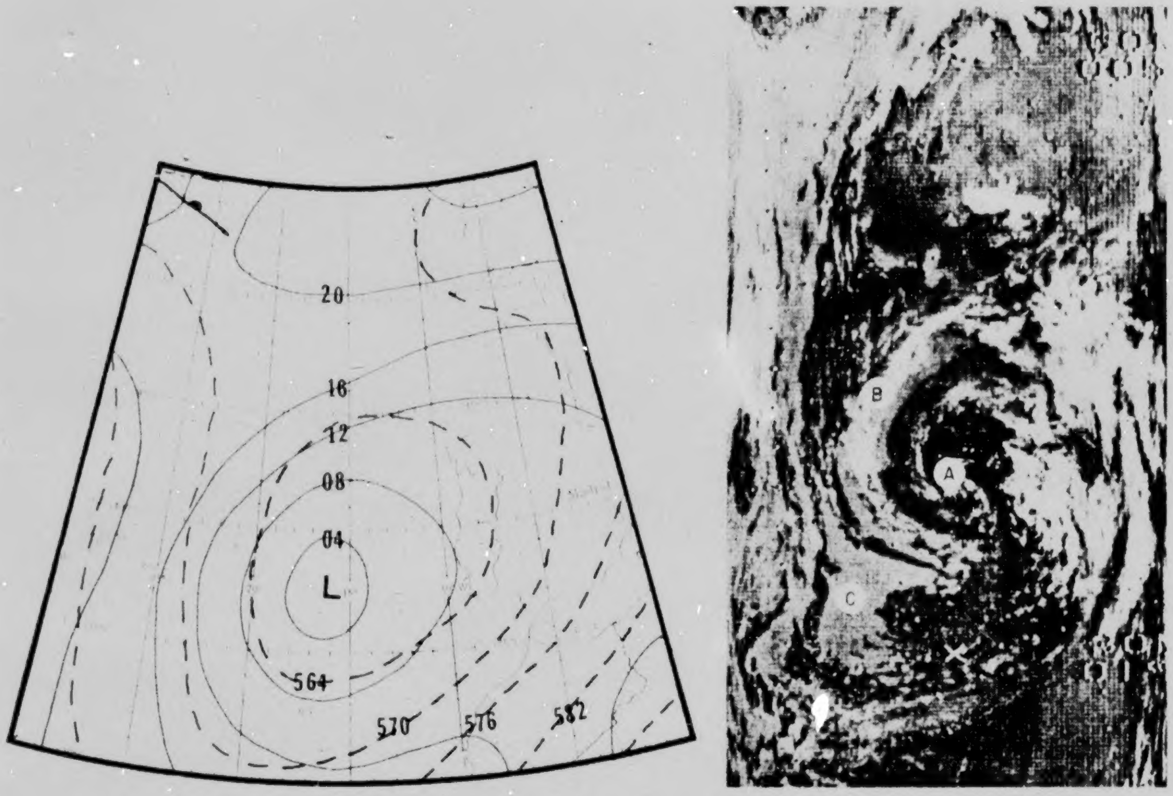
The HRIR picture shows the familiar cloud pattern (fig. 7-4(A)) associated with a cutoff low in the eastern North Atlantic. In this case the cloud bands which spiral in toward the center contain some cumulus congestus, but there is no distinct spiral of low-level clouds which would define the location of the center on a surface map. In fact, there is some question whether or not a surface low center exists.

The general pattern is quite similar in the visual and IR pictures, but the areas of strong vertical motion and active weather are more clearly depicted in the temperature field. For example, the cloud band (fig. 7-4(B) and 7-5(B)), appeared quite bright and solid in the visual pictures taken at both 12 hours before and 12 hours after the IR, whereas the IR showed only small scattered areas of activity. The area shown in figure 7-5(C) appears quite active in the visual mode, but the IR shows that the cloud tops are low and flat.

The preceding comparisons are intended to point out a few of the interpretations which might be derived from the use of satellite pictures. It should be borne in mind that the HRIR nighttime view also yields definitive information for flight briefing purposes regarding icing, turbulence, and type of precipitation in the vicinity of the systems appearing in the pictures.

From the preceding discussion it becomes readily apparent that a comparison of the analysis with the satellite pictures is quite beneficial to the analyst. He can not only verify the existence, intensity, and position of weather systems, but in sparse data areas this may be his only means of completing the analysis.

Information related to the interpretation of satellite pictures as they pertain to map analysis may be found in the Direct Transmission System Users Guide, available from NOAA and Guide for Observing the Environment with Satellite Infrared Imagery, NWRF F-0970-158. In addition, other publications are listed in NavSup 2002.



SURFACE (SOLID) AND 500-MB (DASHED) ANALYSES 0000Z 9 JUNE 1969

IR, LOCAL MIDNIGHT (0118Z 9 JUNE 1969)

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Figure 7-4.—500-mb analysis and accompanying HRIR satellite picture.

REVIEW OF PARAMETERS

Contours

Contours are lines joining equal heights of a given isobaric surface. These lines are usually drawn for 60 or 120 meter intervals. Occasionally they may be drawn for intermediate 30 or 60 meter intervals where the gradient is relatively flat. Wind observations are the primary source of data, along with height values, for drawing contours. That is, the contours are directed and spaced in accordance with observed and calculated winds.

Thickness

The geometeric thickness of a layer of air between two isobaric surfaces is evaluated from the mean virtual temperature of the layer. The thickness may be found by taking the difference between the heights reported at mandatory isobaric levels at each station.

Wind Shear

The wind shear may be either vertical or horizontal and is important in determining turbulence, location of frontal zones, and other factors of meteorological importance.

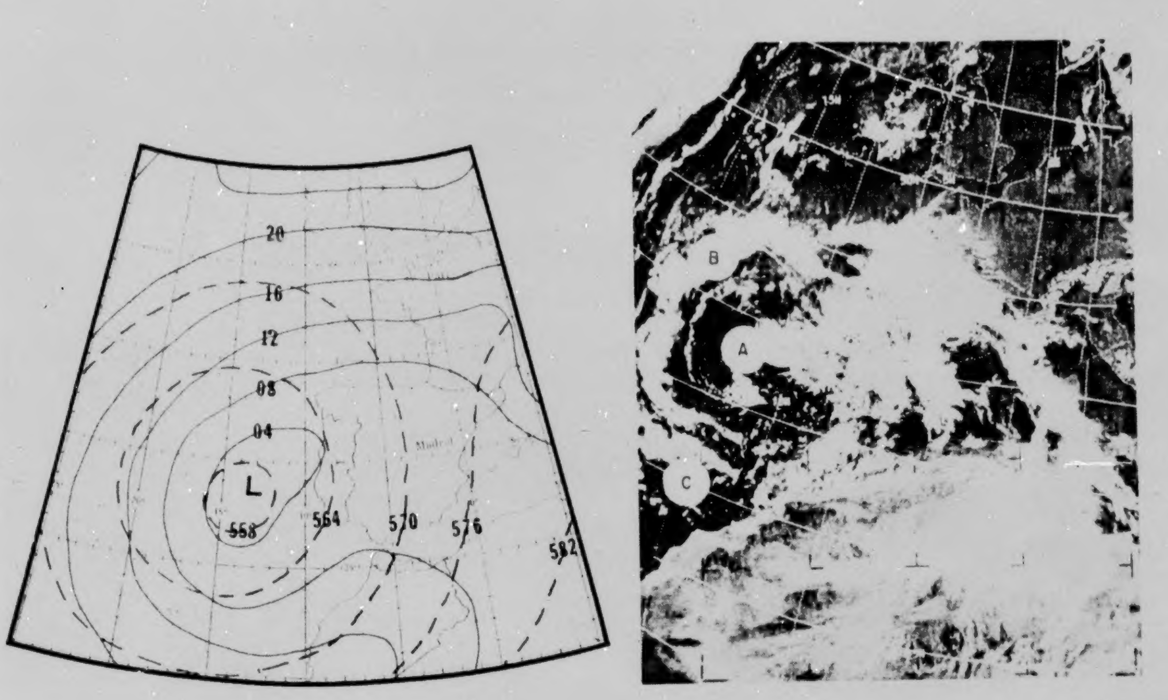

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Figure 7-5.—500-mb analysis and accompanying AVCS satellite picture.

Deepening and Filling

Deepening and filling are terms used to describe processes wherein the pressure at the center of a low pressure system is decreasing with time (deepening) or the system is dissipating (filling). Deepening usually indicates a continued intensification of the gradients of pressure systems.

Strengthening and Weakening

A pressure system is said to strengthen when the gradients and the associated winds become stronger. When the contour gradient becomes flatter or less intense with time, the system is said to weaken.

CONSIDERATION OF CONTINUITY

One basic consideration in the approach to all types of map analysis is that of history or continuity. The slope of height fields and their orientations do not change radically in short periods of time. Consequently, a valuable aid in contour analysis of a given pressure surface is the past history of the contours at that surface. The study of previous maps is essential, both as a key to the present situation, and as a method of preserving an orderly progression of movement and change of atmospheric systems. The necessity for maintaining this continuity is essential for analysis as well as for prognostic purposes.

COMPARISON WITH COMPUTER ANALYSIS

When analyzing upper air charts it must be emphasized that all available data should be utilized. A valuable and often used aid for determining the validity of the manually prepared analysis is the computer analysis. In the not too distant future, as hardware continues to become more available, manually performed analysis will no doubt become entirely outmoded. However, the analyst or forecaster who understands the concepts involved in accomplishing the manual analysis will be able to more intelligently utilize the computer product. Remember, a machine cannot exercise judgement. It must be provided with accurate data and directions for making decisions. Although in most cases the computer product will be more accurate than the manually produced one, the analyst must apply his own judgement based on his experience and other available data. He must not blindly follow any single source, unless this is all he has available. A relatively accurate analysis will be accomplished if comparison is made between satellite data, computer products, and the locally prepared manual analysis (assuming adequate personnel are available).

Types of Computer Products Available

There are over 500 products currently listed in the General Environmental Computer Products Catalog. This publication is published and revised periodically by FNWC Monterey.

In some instances a computer chart may be received which is completely unfamiliar to the recipient. By using the computer products catalog, a description of the chart and its content can be obtained. For example, the GG THETA ANAL shown in figure 7-6 or the CCAT ANALYSIS shown in figure 7-7 might be received at a remote site without explanation.

The catalog would reveal that these charts were an Atmospheric Fronts Analysis and Clear Air Turbulence Analysis respectively.

Procedure for Product Application

The following is a recommended procedure for utilizing the computer products as described in the Computer Products Manual, NavAir 50-1G-522.

1. Determine the accuracy of the "principal analyses" in and upstream of the operational area of concern.

The surface pressure and the 500 mb height analyses are the base upon which all other atmospheric analyses are constructed. Figures 7-8 and 7-9 show examples of these two important charts.

The accuracy of these "principal analyses" profoundly affects the validity of most meteorological and oceanographic prognoses produced within the computer program. Before using any prognostic chart the operational forecaster should satisfy himself that these principal analyses are reasonably correct.

2. If necessary, modify the prognoses which will be used in preparing required forecasts by:

   a. Correcting for any errors detected in the principal analyses.

   b. Correcting for known weaknesses inherent in the prognostic model as discussed previously in chapter 6 of this manual. NOTE: The overhead projector or light table is a valuable aid in accomplishing these two steps.

3. Interpret relevant analyses and prognoses for required meteorological and/or oceanographic forecasts, correcting for known local effects. NOTE: The Local Area Forecasters Handbook, the experience of the forecaster, or textbooks providing data on the area in question may all be brought into play during this phase of the analysis.

SEA LEVEL PRESSURE ANALYSIS.–The sea level pressure analysis was discussed briefly in chapter 6 of this manual. To obtain this chart, computations involving station pressure, present weather, visibility, and ceiling are extracted from each report. Objective analysis of station pressure values is performed using the previous analyses as described in chapter 6. The program output may either provide a hemispheric product as shown in figure 7-8 or, with a variation in the gross error check and the analyses used, a surface pressure analysis of the United States may be obtained. On the analysis limited to the United States more detailed information is shown. This includes significant weather symbols and station flight conditions. For an illustration of this chart refer to chapter 3 of the Computer Products Manual, NavAir 50-1G-522.

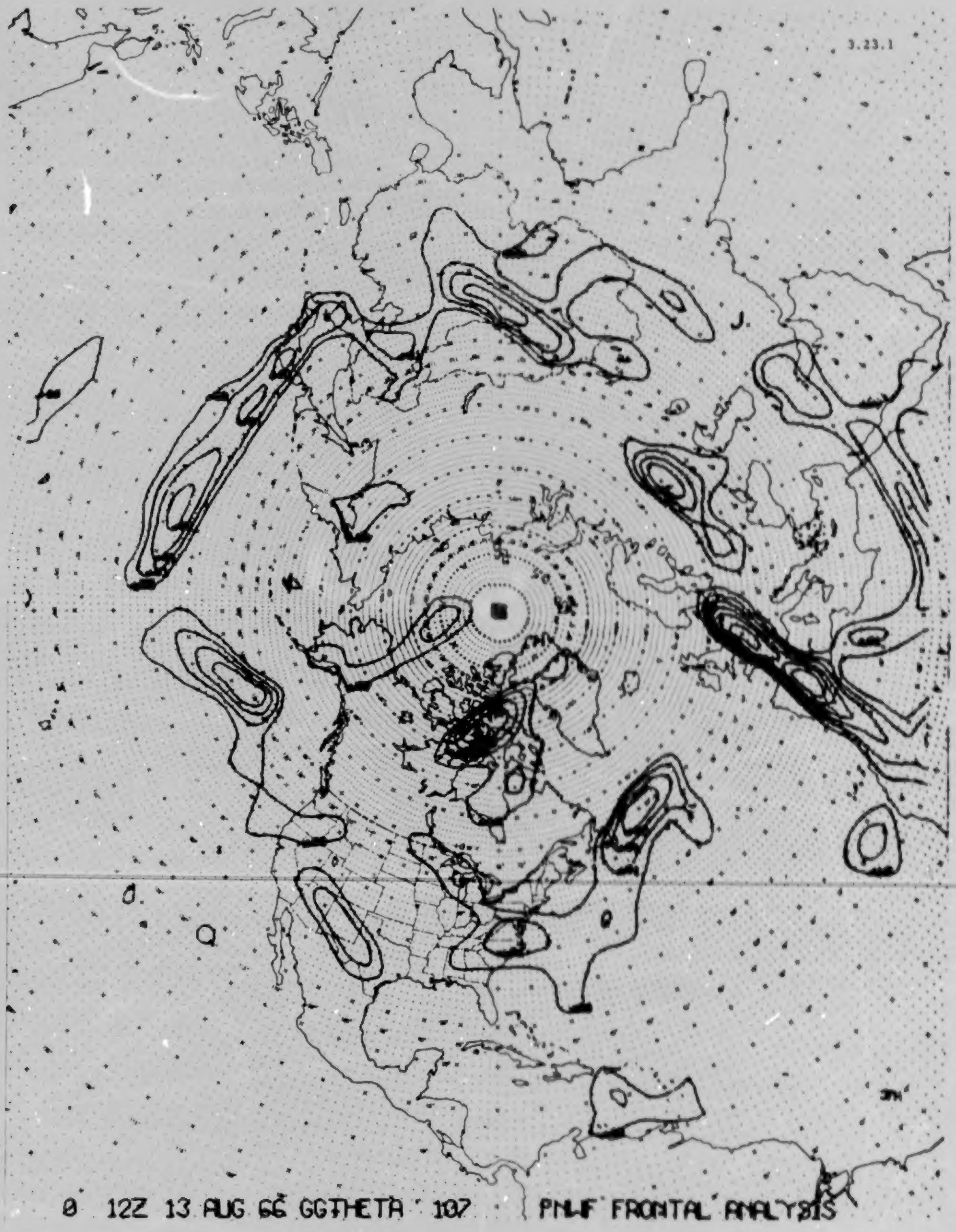
3.23.1
0 12Z 13 AUG 66 GGTHETA 107 PNWF FRONTAL ANALYSIS

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Figure 7-6.—Atmospheric fronts 36-hour prognosis. Axis of patterns depicts warm boundary of strong 1000-mb temperature gradients. Strength of temperature gradient is proportional to number of isolines surrounding pattern axis.

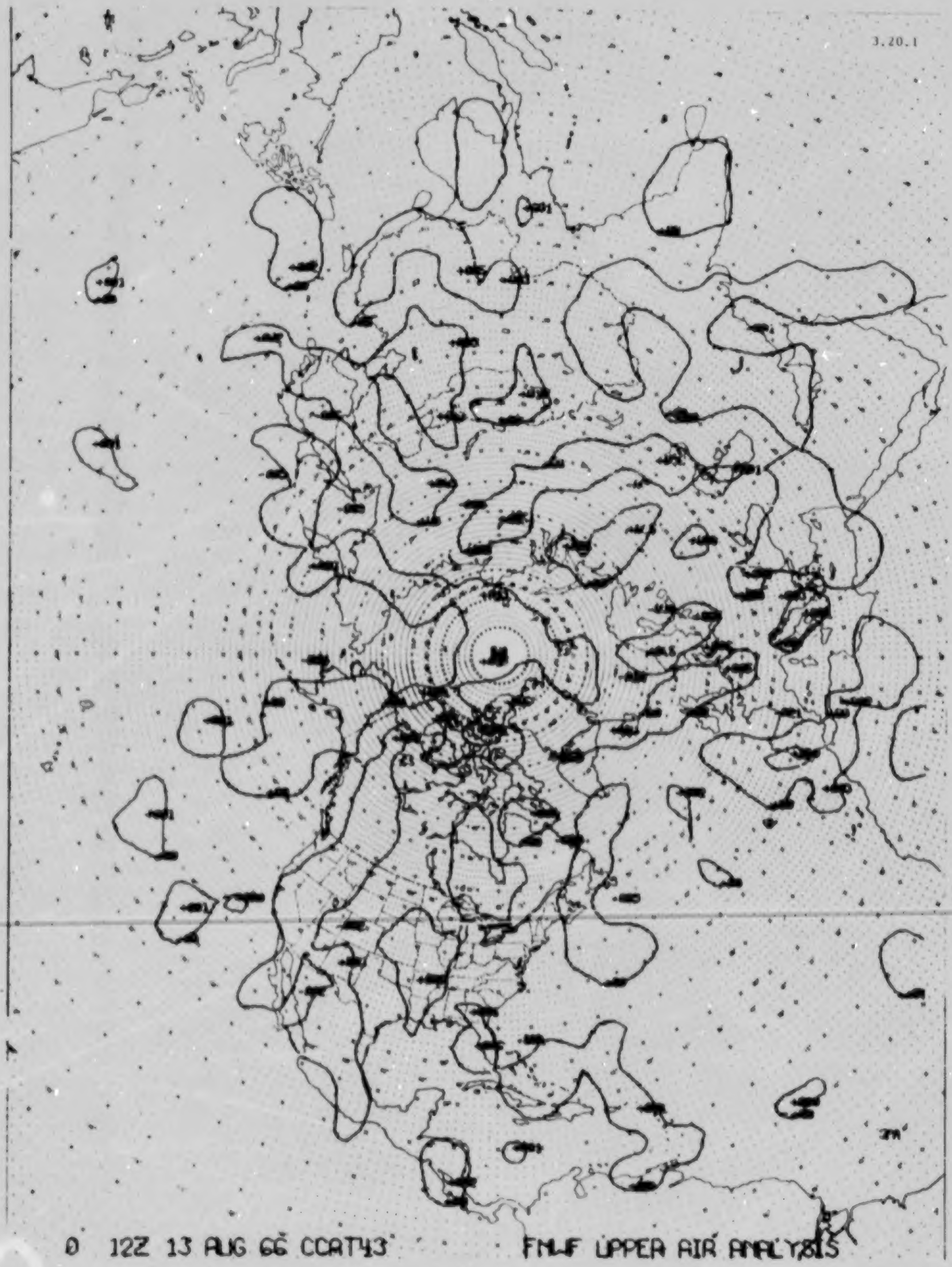
3.20.1
0 12Z 13 AUG 66 CCAT43
FNWF UPPER AIR ANALYSIS

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Figure 7-7.—Clear air turbulence in the 400- 300-mb layer analysis. Contour interval 25 relative units.

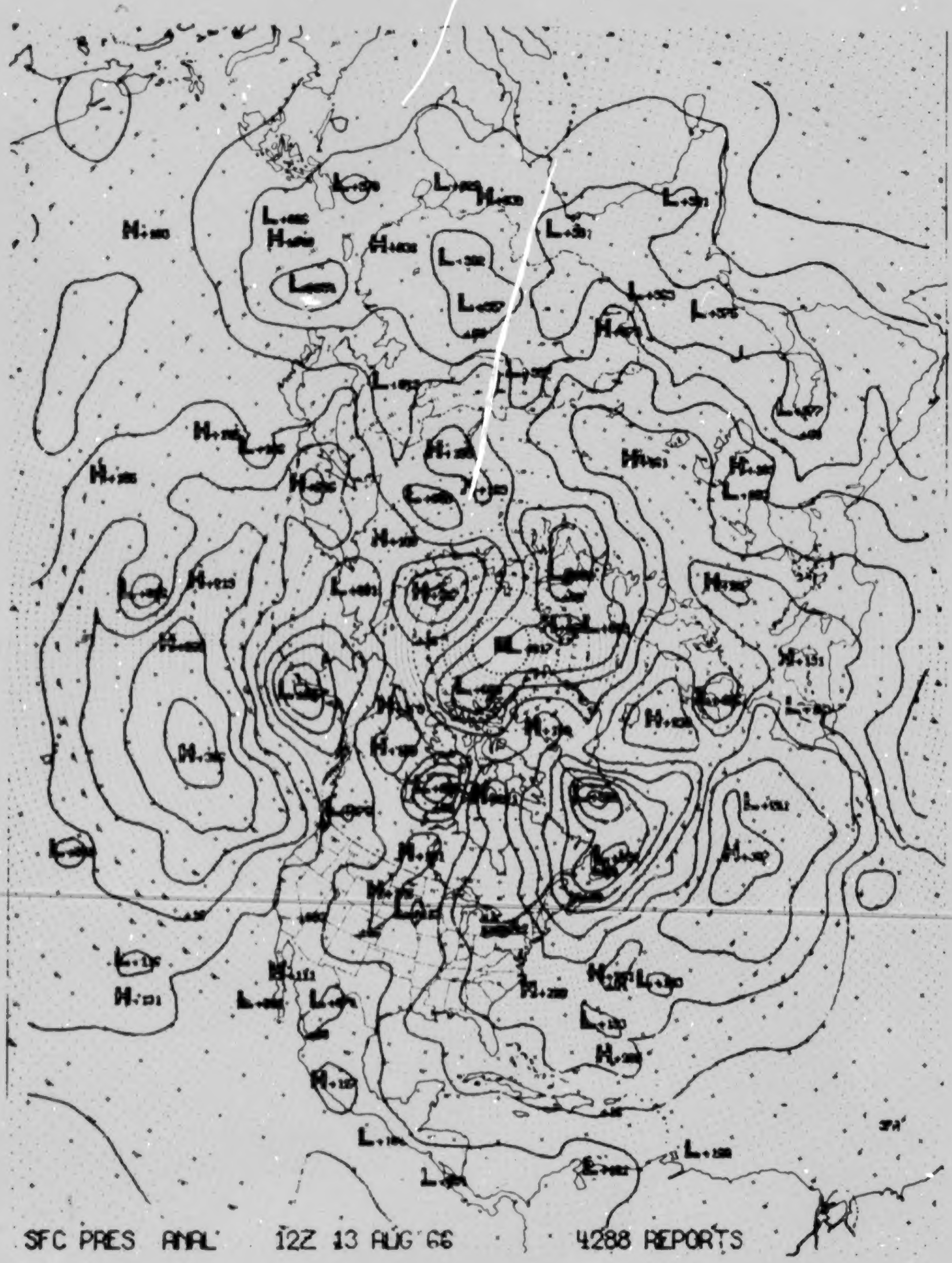
SFC PRES ANAL 12Z 13 AUG 66 4288 REPORTS

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Figure 7-8.—Sea level pressure analysis. Contour interval 4 mbs.

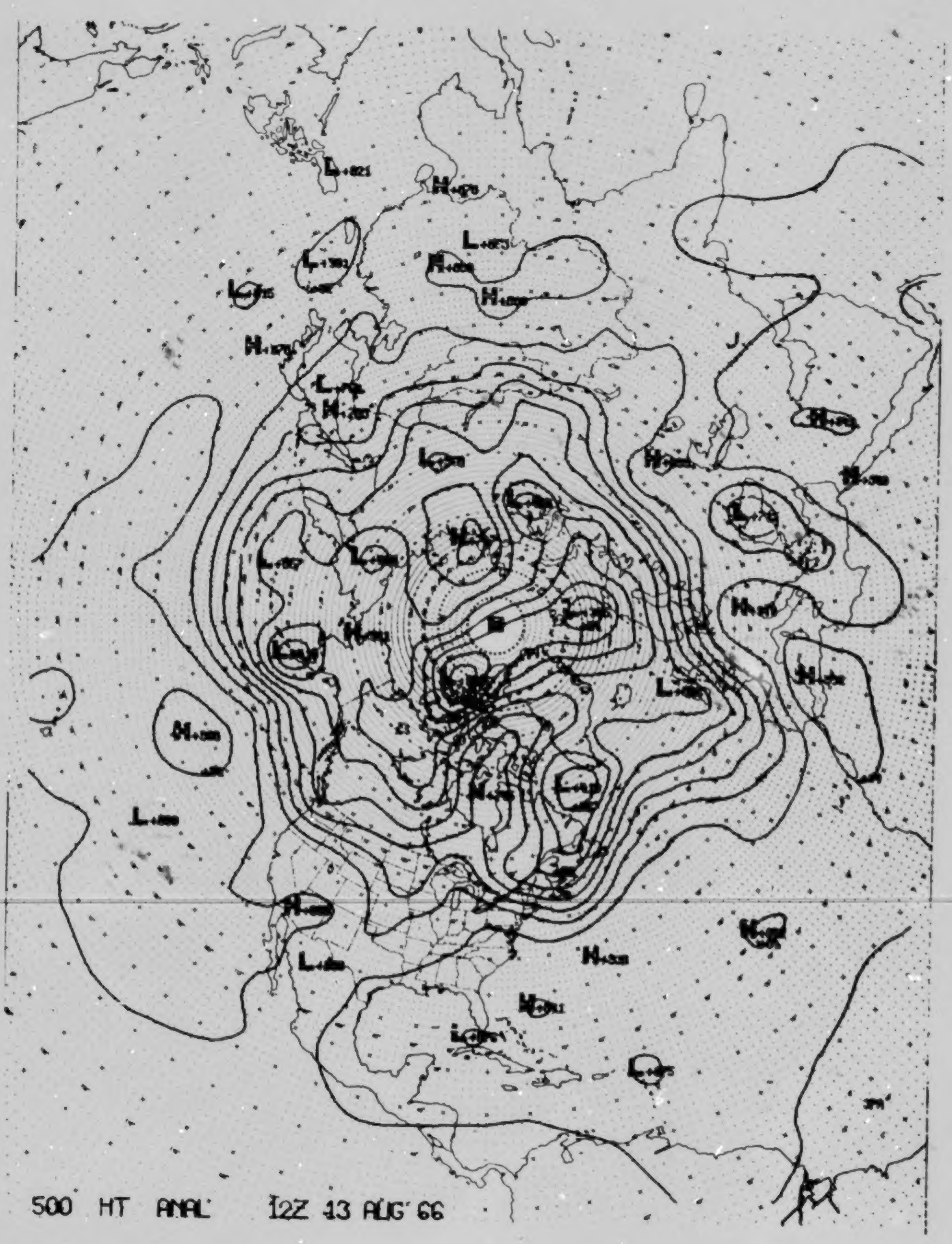
500 HT ANAL 12Z 13 AUG 66

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Figure 7-9.—500 mb height analysis. Contour interval 60 meters.

500 MB ANALYSIS. The 500 mb analysis shown in figure 7-9 is obtained through the following procedure:

1. Objective analysis of reported 500-mb heights and heights extrapolated by the geostrophic approximation from reported 500-mb winds is performed. The first guess is obtained by reanalyzing the 12 hour previous 500-mb analysis using all data collected during the 12 hour period between synoptic observation times. The "update" analysis usually contains 10 to 20 percent more 500-mb data than did the original. From the update, a 12 hour 500-mb prognosis verifying at the present map time is derived. This prog is then modified for the 12 hour change in 500- to 1000-mb thickness inferred from the present surface pressure analysis. The prog is also adjusted toward climatology in regions where the update analysis was constructed from sparse data.

2. Three cycles through the analysis routine are made with the data. The first and second passes with data utilize only the height values from the reports. The second pass provides lateral checking wherein the datum value must be within a prescribed tolerance of the value interpolated from surrounding data.

3. Upon completion of the second pass, the reported winds are used to compute geostrophic gradients and derived height values at four surround points. The extrapolations parallel to the wind are placed 1.5 mesh lengths from the datum and the normal extrapolations are placed 0.75 mesh lengths away. (South of 30N these distances are changed to 2.0 and 1.0 mesh lengths respectively). If another report of height and wind falls within this same area, the distances will be modified so that the extrapolations are located no closer than half the distance between the two stations. In the case of aireps with no height values reported, an interpolation of the analysis is used to provide the working height values.

4. The third and last analysis pass is made using the heights and wind extrapolations with lateral checking as defined previously. The final analysis field is then modified where necessary so that the absolute vorticity always exceeds one-third of the coriolis parameter.

SMOOTHING OF FLOW AND GRADIENTS

It should be kept in mind that the atmospheric circulations of interest are of a relatively large scale. Consequently, the field of flow represented by the contour lines should normally be a smoothly continuous field. However, contour gradients are seldom uniform or constant over large areas. Rather, at places such as near jetstreams and near fronts, both contour and thickness lines in the lower troposphere are often oriented somewhat parallel to these phenomena and their spacing is considerably closer than at other areas.

Small-scale perturbations which are not apparent in two or more contours and which cannot be associated with some surface system should usually be smoothed out of the 700- and 500-mb analyses. On the other hand, it is erroneous to assume that gradients remain constant over large areas. Nevertheless, unless there is definite evidence of discontinuous conditions or important small-scale phenomena, the analysis should reflect a uniform progression of flow and its time changes. Such a smooth analysis will usually be more correct than if sporadic changes are indicated. In this aspect, you should use the classical models of fronts, jetstreams, etc., to aid in depicting the most likely distribution of winds and contours in regions of sparse data.

For example, it is the policy of the National Meteorological Center to smooth out disturbances on surface and upper-level charts of wavelengths less than about 8 degrees latitude. This policy is based primarily on the belief that the majority of disturbances of wavelengths less than 8 degrees latitude are fictitious and the result of inaccurate observations and are nonrepresentative of the large scale features.

HISTORICAL SEQUENCE

Historical sequence is vitally important to good map analysis. Some of the reasons for this have been pointed out in a previous section titled "Consideration of Continuity." Consequently, one of the first steps in the analysis should be to check the previous charts for accuracy, rationality, and any corrections made subsequent to the initial analysis. Fronts, troughs, and pressure centers do not normally appear or disappear from one chart to another and any such occurrence should be viewed with suspicion.

The past positions of all pressure centers should be entered on the current chart for a period of at least 24 hours and for longer periods when practicable. These positions are normally entered in black ink with an X circumscribed with a circle and connected with a dashed line. The time and data are entered above the circle. The corrected positions of all fronts, troughs, and ridges should be transposed on the current chart in yellow pencil. Do not forget the short wave troughs. These trough move around and through long wave troughs and are often important in the genesis of fronts and cyclones.

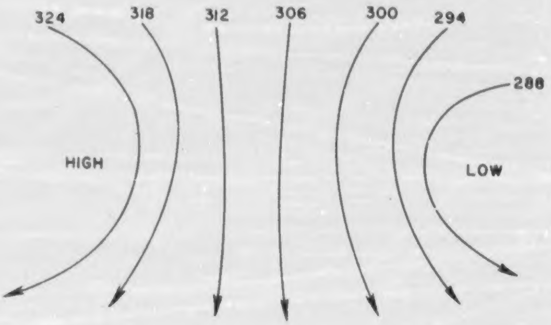
CONTOUR ANALYSIS

The rules given in this section generally apply to all upper level charts. The analysis of the jetstream by isotach analysis is covered in a later section. The primary intervals of contour spacing are 60 and 120 meters. The 60-meter interval is used on charts from the surface up to 300 mb and the 120-meter interval is used on charts at 300 mb and above. A solid black line is used to represent these primary contours. Intermediate contours may be used when greater definition is needed: a 30-meter interval is used with the 60-meter primary interval and a 60-meter interval is used with the 120-meter primary interval. Intermediate contours are represented by a dashed black line.

Before beginning the contour analysis you should first compute heights in areas of sparse data, using the principles outlined in a previous section of this chapter. Such approximations are most reliable over ocean areas. Plot these approximations on the chart and enclose them in parenthesis.

RULES FOR DRAWING CONTOURS (NORTHERN HEMISPHERE)

Except where reports are closely spaced, there are usually many different sets of contours that can be drawn to a given distribution of height reports; however, only one set is correct. Therefore, lines drawn merely to data do not represent an analysis. A reasonably correct analysis can be obtained only by applying careful judgement to all factors involved in the analysis. The following discussion of rules for drawing contours in the Northern Hemisphere is illustrated by figures 7-10 and 7-11.



AG.524

Figure 7-10.—Contour pattern between a high and a low (700-mb heights in meters).

1. Between adjacent highs and lows, there will never exist two contours of the same value as illustrated in figure 7-10.
2. Between two adjacent highs there will always be two contours with the same value (A and B in fig. 7-11).

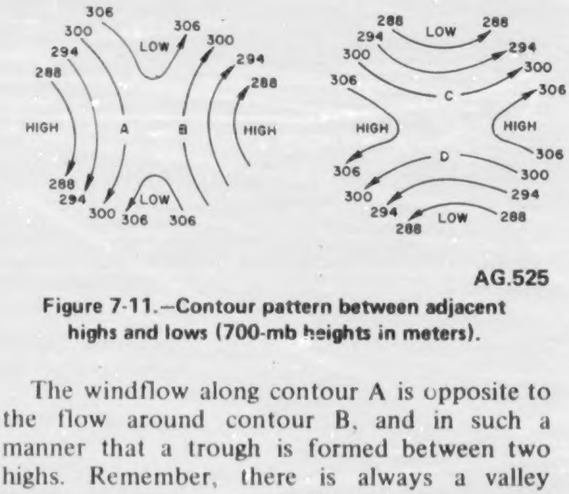


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Figure 7-11.—Contour pattern between adjacent highs and lows (700-mb heights in meters).

The windflow along contour A is opposite to the flow around contour B, and in such a manner that a trough is formed between two highs. Remember, there is always a valley between two mountains. The configuration of contour A is independent of the shape of contour B. The height between A and B is less than the value of A or B, but greater than the value of the next lower contour. The region of lightest winds usually exists between A and B.

3. Between two adjacent lows there are always two contours of the same value (C and D in fig. 7-11). The mean windflow along contour C is opposite to the flow around contour D and in such a manner that a ridge is formed between two lows. Remember, there is always a mountain between two valleys. The configuration of contour C is independent of the shape of contour D. The height between C and D is greater than the value of C and D, but less than the value of the next highest contour. The region of lightest winds usually exists between C and D.

COMMON ERRORS IN ANALYSIS

An analyst, especially an inexperienced one, will have difficulty drawing contours in areas of sparse reports. Figure 7-12 illustrates some typical errors in contour analysis.

SKETCHING CONTOURS

Contours should first be sketched lightly, using black pencil in accordance with the reported and computed heights and winds. As stated previously, contours are parallel to the wind direction and spaced inversely proportional to the wind speed. Geostrophic wind scales should always be kept near at hand. These scales have been discussed previously in this chapter and in chapters 4 and 6. These scales should be used as an aid in spacing contours, especially in areas of sparse data. Some precautionary rules should be kept in mind when using these scales. For stationary troughs and ridges, cyclonically curved contours have a closer spacing of contours for the same speed when the contours are anticyclonic. Too, in regions of subgradient or supergradient winds, the winds blow across contours. The contours normally show a slight kink on fronts at the 850-mb level, but show up as troughs above this level.

The sketched contour pattern is adjusted for reported winds and maintenance of continuity. The pencil used to draw contours should be carefully selected so that the sketching can be cleanly and easily erased. A medium lead pencil is recommended.

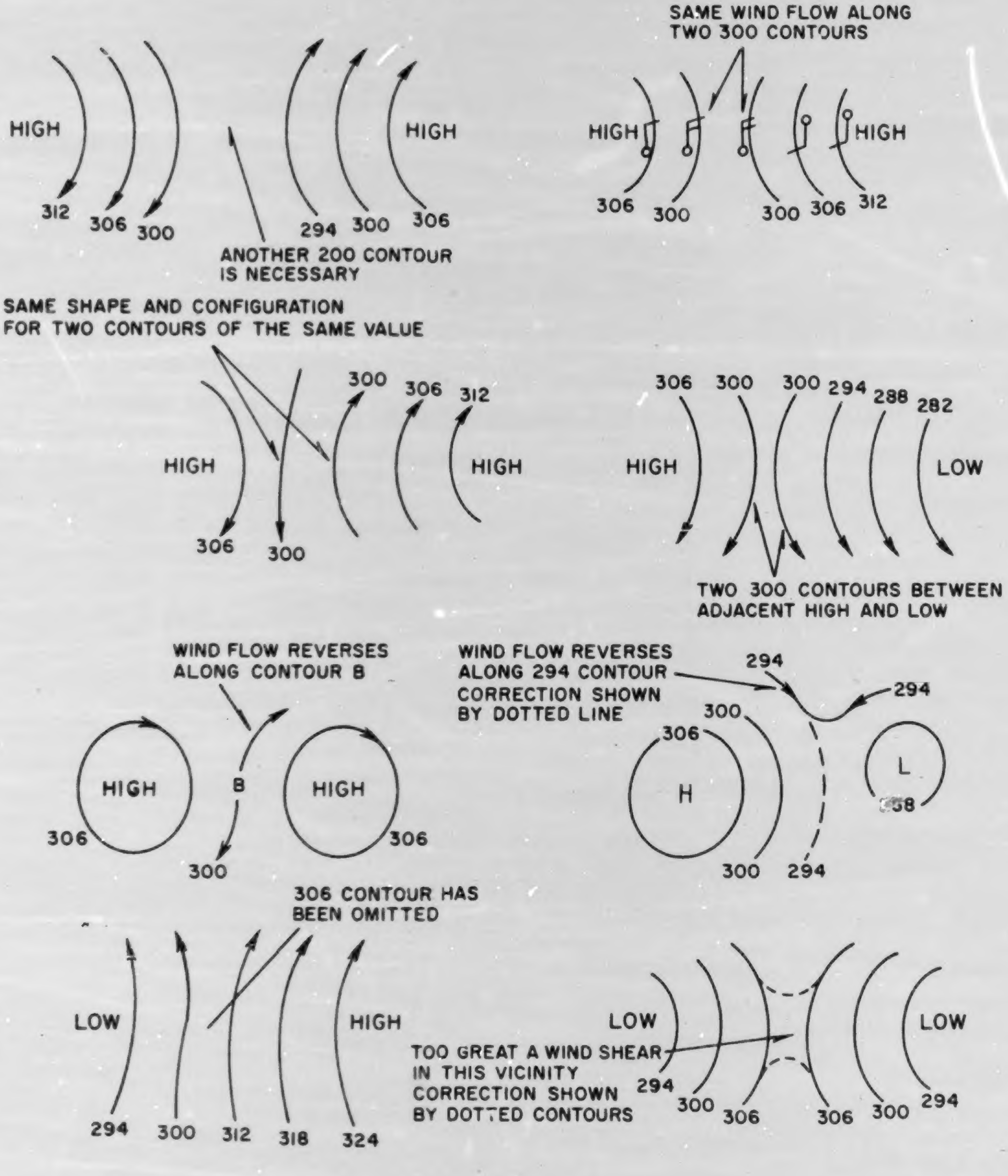
Contour labels consist of numbers representing the whole value of the line in meters for which each particular contour is drawn. For example, the 3060 meter contour on the 700-mb chart would be labeled 306, dropping the tens digit. An open contour is labeled at both ends; a closed contour should be broken at one convenient spot to permit entry of the label, usually at the northernmost position of the particular contour line. The labels for a series of closed, concentric contours should be arranged to form an easily read line of numbers running from low to high contour values. All contour labels should be of uniform size, their bases should be parallel to the adjacent circles of latitude, and they should be in the same color and with the same pencil used to draw the contour lines. Labels should be neatly printed.

In sketching preliminary contours, do not try to manipulate the pencil with just your fingers, but make smooth sweeping lines, bringing your entire arm into motion. Have your hand in such a position that it does not obstruct the view of the reports with which you are immediately concerned. Keep your eyes just ahead of the pencil, thereby determining the points through which your contour should pass. This will enable you to anticipate changes in direction which the contour should take so that these changes may be smooth rather than abrupt.

All contours are continuous lines, closing either within the limits of a particular chart or beyond the margin of the chart being drawn. In no cases do contours cross other contour lines, join contour lines of different values, or become broken.

Wind direction and wind speed are of primary importance in drawing contour lines. Each contour line should be drawn parallel to nearby winds whose speeds are 10 knots or greater. While constructing the contour pattern, the analyst should continually strive to represent the wind field accurately.

Allowance must also be made for the possibility of errors in the data when drawing contours. The experienced analyst must constantly bear in mind the factors which contribute to the variety of errors. In general, it is proper to neglect


HIGH
312
306
300
ANOTHER 200 CONTOUR IS NECESSARY
294
300
306
HIGH
SAME WIND FLOW ALONG TWO 300 CONTOURS
HIGH
306
300
300
306
312
HIGH
SAME SHAPE AND CONFIGURATION FOR TWO CONTOURS OF THE SAME VALUE
HIGH
306
300
300
306
312
HIGH
306
300
300
294
288
282
HIGH
LOW
TWO 300 CONTOURS BETWEEN ADJACENT HIGH AND LOW
WIND FLOW REVERSES ALONG CONTOUR B
HIGH
306
B
300
HIGH
306
WIND FLOW REVERSES ALONG 294 CONTOUR
CORRECTION SHOWN BY DOTTED LINE
294
294
300
306
H
300
294
L
306 CONTOUR HAS BEEN OMITTED
LOW
HIGH
294
300
312
318
324
LOW
LOW
TOO GREAT A WIND SHEAR IN THIS VICINITY
CORRECTION SHOWN BY DOTTED CONTOURS
294
300
306
306
300
294

AG.526

Figure 7-12.—Common errors in contour analysis (700-mb chart).

unusual departures from the overall height distribution and draw each contour as a smooth curve, omitting irregularities that would exist if the contours were drawn to the absolute value of each individual height value reported. Normally, however, the lines should not be smoothed if the irregularities are indicated by two or more adjacent contours. When reports are sparse, however, no report should be disregarded unless thoroughly checked.

Application of Satellite Cloud Photographs

The application of satellite cloud photographs to the analysis of contours is somewhat limited and dependent to a considerable degree upon the experience of the analyst. However, there are certain aspects which are helpful. For example, the orientation of troughs and ridges may be verified by the existence of certain cloud features. Figures 7-4 and 7-5 illustrated to some degree the contour pattern vs satellite presentation about a cutoff low.

Figure 7-13 depicts an HRIR picture of a frontal band and associated 300-mb short wave trough in the eastern Pacific. Here the highest clouds, which radiated at the coldest temperature, appear the whitest, and the lower warmer clouds appear dark gray. In this case, the high bright clouds begin abruptly at 0, just east of the point where the 300 millibar trough (dashed line) intersects the frontal band.

To the west of the trough line, the frontal band (Q) appears dark gray; here the frontal clouds are warmer and lower than the clouds east of the trough. The darkest areas, (P) correspond to clear sky views of the warm waters of the Pacific Ocean. The contour gradient is indicated to some degree by the spacing of the cloud features.

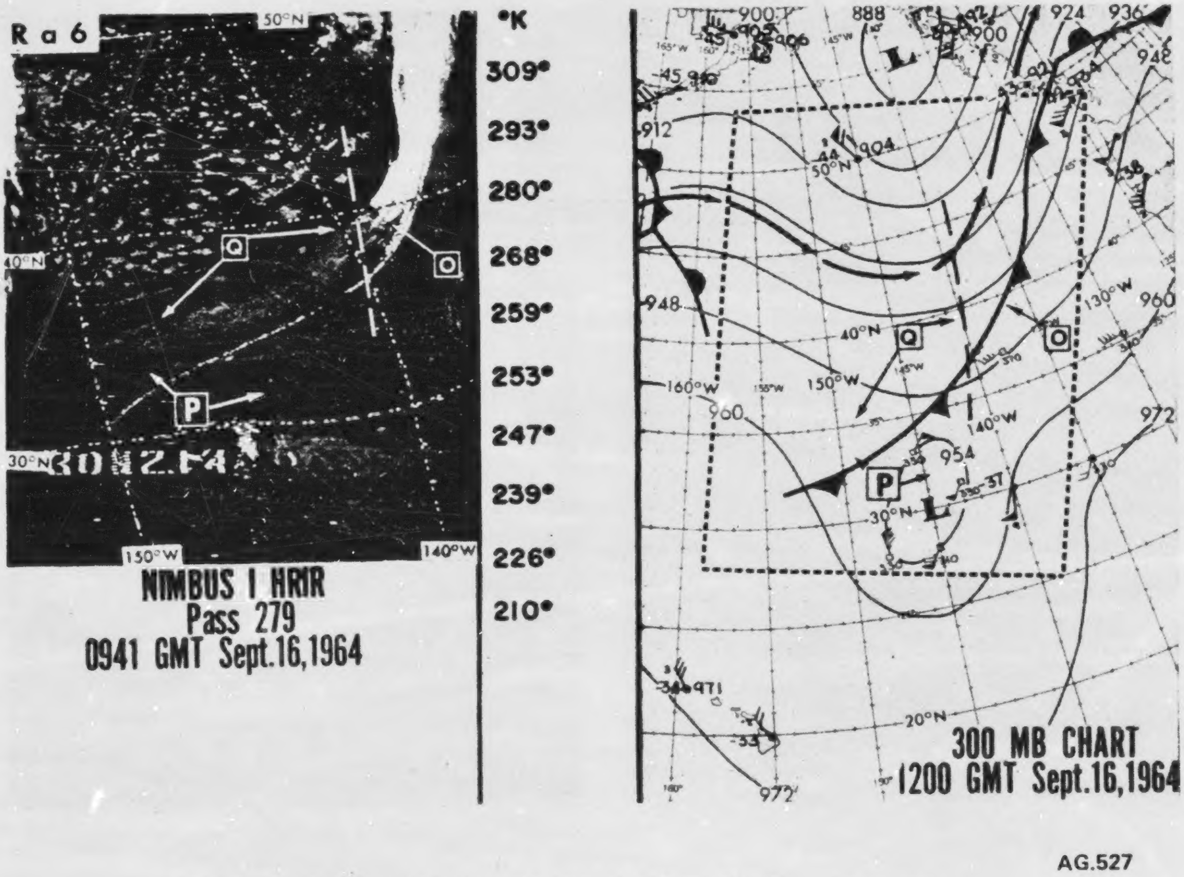

R a 6
50°N
40°N
30°N
150°W
140°W
Q
O
P
°K
309°
293°
280°
268°
259°
253°
247°
239°
226°
210°
NIMBUS I HRIR
Pass 279
0941 GMT Sept.16,1964
L
50°N
40°N
30°N
20°N
160°W
150°W
140°W
130°W
912
948
960
972
954
300 MB CHART
1200 GMT Sept.16,1964

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Figure 7-13.—Evidence of an upper level trough and associated contour pattern in HRIR data and accompanying 300-mb analysis.

The position of an upper level ridge and associated contour pattern are illustrated in figure 7-14.

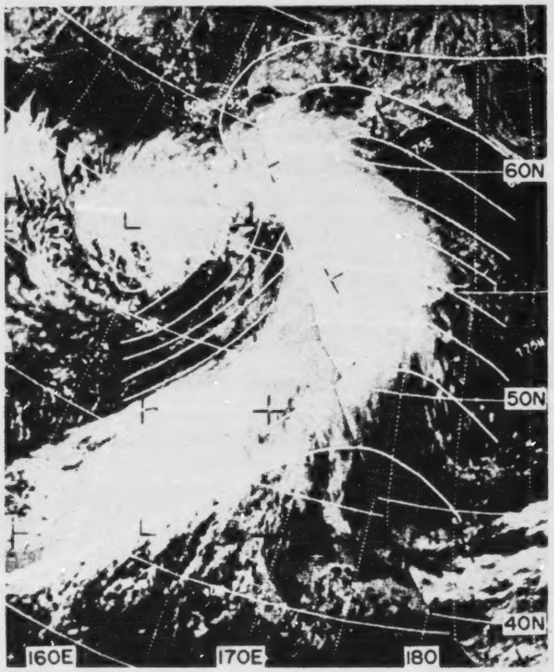

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Figure 7-14.—500-mb flow pattern superimposed on AVCS satellite photograph showing the position of an upper level ridge.

The accuracy with which the position of a ridge line can be located is strongly affected by the amplitude of the ridge, and its wind field, which determine the associated cloud structure. If the upper level ridge has sharp anticyclonic curvature, the multilayered frontal clouds dissipate rapidly in the area of subsidence downstream from the point where the upper level flow has its maximum anticyclonic curvature. Notice that in figure 7-14 the eastern edge of the frontal cloudiness ends abruptly at the 500-mb ridge. The solid lines show the 500-mb flow pattern. If the wind is very strong or the anticyclonic curvature of the upper level ridge is quite gradual, the multilayered clouds are more diffuse in appearance east of the ridge. Under either of these conditions, the position of the upper level ridge is less clearly defined by the clouds.

APPLICATION OF COMPUTER PRODUCTS

The application of computer products to contour analysis has been found to be of such value that in the majority of instances the computer product is used in place of the hand drawn analysis. However, as mentioned earlier in this chapter, an understanding of the concepts utilized in the hand drawn analysis will enhance the analyst's or forecaster's ability to utilize the computer product.

Given correct data in sufficient density the computer performs analyses with an accuracy exceeding that of the instruments used to measure the elements being analyzed. Computation of winds or pressure-height values using the geostrophic wind equation is performed wherever desirable. The computer is capable of handling complex interactions with ease and eliminating unreasonable noise and stability values. Continuity in space and time is maintained to a degree beyond the capability of hand methods. The computer is wholly objective, and eliminates individual differences in the analysis. If the program and the data are correct, the computer product will be correct.

The recommended procedure for applying the computer product to the hand drawn analysis was presented earlier in this chapter.

There are many computer charts available to assist in the location of ridge and trough lines. These charts are listed in the Computer Products Manual, NavAir 50-1G-522. The procedures to be followed and considerations to be made in their application were mentioned in chapter 6 when discussing surface analysis and earlier in this chapter when discussing hand drawn upper air analysis. Computer prognoses are excellent aids for determining future movement of these features and will be discussed in chapter 8 of this manual.

FINISHING CHART

Finally, after a reasonably accurate contour pattern has been analyzed, taking into consideration the factors mentioned above, the sketched contours should be erased and smooth flowing lines, except where noted above, drawn on the chart. Then the contour labels should be placed in their final form in accordance with previous instructions. The troughs, ridges, highs, and lows should be indicated by the appropriate symbols.

CONTOURS IN RELATION TO PRESSURE SYSTEMS

As pointed out in chapter 4 of this training manual, the most common upper air pattern consists of alternate troughs and ridges, with occasional closed lows in the troughs and closed highs in the ridges. This is illustrated in figure 7-15.

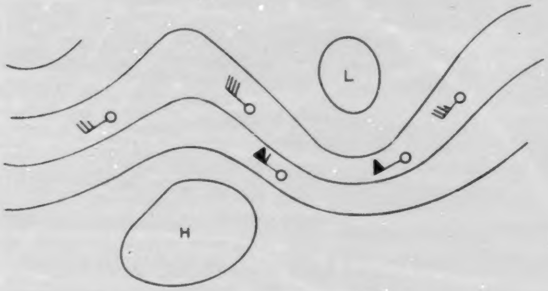

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Figure 7-15.–Common upper air patterns.

These troughs and ridges form a series of waves which encircle each hemisphere with lows normally placed poleward and highs equatorward. These lows and troughs are generally associated with cyclones at lower levels, sea level in particular.

Similarly, ridges and highs often reflect the vertical extent of surface anticyclones. (See fig. 7-16.) Note that the axes of the systems shown in figure 7-16 are not vertical; the axis of the cyclone or trough always slopes upward toward colder air (usually westward and poleward) and the axis of the anticyclone or ridge always slopes toward the warmest air (usually westward and equatorward). These spatial relations are an absolute requirement for the proper 3-dimensional representation of pressure systems.

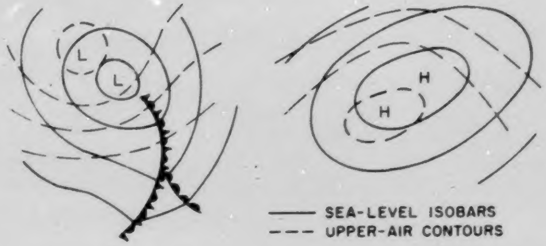

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Figure 7-16.–Slope of surface pressure systems with height.

Other closed lows sometimes appear on upper air charts and work down to the surface. These systems have relatively cold cores and are called cold core lows. High level anticyclones not associated with surface highs are much rarer and never attain an intensity comparable to that of cold lows.

CONTOURS IN RELATION TO FRONTS

Except on the 1,000-mb chart and along cold fronts on the 850-mb chart, contours kink only slightly or not at all, in crossing fronts. Above 700 mb, fronts are usually omitted from the analysis. It is unusual to find the upper portions of fronts above 850 mb located along upper trough lines, except for very short periods of time. Usually the upper portions of cold fronts are located somewhere in advance of the upper trough, with very slight shifts of wind across the frontal zone and almost impreceptible kinks in the contours. The major wind shifts and contour troughs occur in the rear of the upper portion of the frontal surface. When upper portions of cold fronts are found along these upper trough lines, the fronts are at maximum intensity. Usually, the upper trough moves much more slowly than the wind, and since the cold front moves with the wind speed, except for the slowing effect of subsidence in the cold mass, the front usually outstrips the trough until it reaches the southwest flow in advance of the trough, where its eastward movement slows down.

Upper portions of warm fronts are seldom ever associated with troughs or perceptible cyclonic kinks in the contours of the constant pressure surfaces above 800 mb, assuming the ground is near sea level. Above about 700 mb the warm front intersection is usually found near the ridge line. Not only are there no apparent cyclonic kinks in the contours crossing the frontal zones, but frequently, contour curvatures and wind shifts appear anticyclonic. At 700 mb the anticyclonic curvature is the rule rather than the exception. At 700 mb and above, no attempt should be made to kink the contours across upper portions of fronts which have been found in the frontal analysis.

ANALYSIS OF SPECIFIC FEATURES

ISOTHERM AND FRONTAL ANALYSIS

Drawing Isotherms

Isotherms (lines of equal temperature) are usually drawn at intervals of 5° C. They are drawn as solid red lines. Isotherms should be sketched in according to reported temperatures, using past isotherm positions as a guide. Along with wind and pressure, isotherms on the constant pressure charts yield clues as to the exact state of the atmosphere at the present time and aids in the construction of prognostic charts.

After sketching the isotherms and checking their consistency, the isotherms should be drawn in as smooth solid red lines, heavy enough to be easily legible, but not so heavy that they detract from the contours. Label the isotherms near the edge of the chart. For those which form closed curves, leave a small break in the curve near the top of the curve for labeling.

Isotherm Patterns

Isotherm patterns at lower levels usually consist of tongues of warm and cold air that move across the map with reasonable continuity from day to day. There is a correlation between the movement of the isotherm pattern and the circulation, and between the form of the isotherm pattern and the distribution of contours. Warm tongues generally tend to coincide with ridges and cold tongues with troughs, particularly at higher levels, such as 500 mb and above, but this is only true as a first approximation. It is actually the departures from this relationship that are most important, and these can be determined only by making full use of all available data and by having a knowledge of the picture normally to be expected in a given situation.

The isotherms must be reasonable dynamically and kinematically. Kinematically, the isotherms should move with the wind if the air parcels conserve their temperature. Usually isotherms move somewhat more slowly than the wind speed, depending upon the amount of vertical motion and the amount of heating or cooling.

Dynamically, the warm tongues in the troposphere are related to pressure ridges, due to sinking and adiabatic warming resulting from high-level convergence. If the ridge attains sufficient amplitude, the northern end may be cut off into a separate high cell at upper levels. This results in a cutoff high. Upper-level convergence also causes ascent of air and cooling in the lower stratosphere at the 300-, 200-, and 100-mb levels and above. Thus, the building of an anticyclone aloft is characterized by an increasing warm tongue in the troposphere and an increasing cold tongue in the lower stratosphere.

Cold tongues are related to pressure troughs in the upper troposphere, due to upper-level divergence, resulting in ascent of air in the lower and middle troposphere, with adiabatic cooling. An additional consequence of divergence in the upper troposphere is sinking in the lower stratosphere accompanied by adiabatic warming. Deepening of a low aloft tends to be associated with an intensifying cold tongue in the troposphere. This is illustrated in some cases of cold lows aloft where the isotherms are parallel to the contours and no advection is indicated. As further deepening occurs, the isotherms around the cold core are seen to move radially outward, across the wind, due to dynamic cooling.

Movement of Isotherms

In regions of sparse data an estimate of the current contour pattern, based on the previous temperature field and circulation integrated with the latest surface pressure pattern, aids greatly in determining the most nearly correct pattern aloft. This is substantially the procedure involved in graphical addition of thickness patterns to lower-level contour patterns to produce consistent upper-level contour patterns. Temperatures at selected points are frequently used in the computation of heights at various upper levels, instead of using the differential method, although the latter is used by some analysts.

Keeping this first approximation in mind as the isotherms are drawn, the Aerographer's Mate finds that even in regions of scarce data, much of the departure from this relationship becomes evident from the data. But, where data are scarce, it is necessary to examine further evidence. In this respect, certain conditions must be considered, and these are discussed in some detail in succeeding paragraphs.

It is important to note that a continuous rise of temperature cannot always be distinguished from a case where the temperature was rising during most of the period between reports, and then began to fall shortly before the time of the current report. Similar reasoning would apply to the passage of a cold tongue with opposite temperature changes being observed.

Where there is a change of wind with height, the isotherm pattern is oriented to conform to the shear vector and the spacing of the isotherms is made to conform with the magnitude of the shear vector. Over land, where there is a dense network, wind shear is not used explicitly in the determination of the isotherms. Over the oceans computed shear vectors are of considerable value in drawing isotherms. Speeds of less than 10 knots may not be significant, but when the shear vector exceeds this value, it is of particular use in isotherm analysis.

The assumption that isotherms in constant pressure surfaces at the middle of an isobaric layer are closely similar to the mean isotherms (thickness lines) of the layer is borne out by a comparison of the 850-mb isotherms with the mean isotherms of the 1,000- to 700-mb layer, and by comparing the 700-mb isotherms with the mean isotherms of the 1,000- to 500-mb layer. The 700-mb isotherms constitute a good advection chart for the layer from 1,000 to 500 mb.

Also see chapter 10 of this training manual for additional information on movement of the 850-mb isotherms.

The thermal winds and reported temperatures are not the only guide to drawing isotherms on upper-level charts. Reasonable continuity in time from chart to chart and a reasonable consistency between the various levels for the same time are sought at all times. It is important in regions of scarce data that the surface chart be examined carefully in order that the main features of the upper-level isothermal pattern will ~~be~~ consistent within reasonable limits with what is expected from the sea level pressure distribution and frontal systems.

After the isotherm analysis is completed, the movement and development of the temperature pattern should explain the temperature changes reported at individual stations. The temperature changes at any station can be explained as the result of:

1. Advection.
2. Changes due to the addition or subtraction of heat, and adiabatic changes due to lifting or subsidence.

Isotherms generally move with a speed somewhat less than the wind speed. Two of the main reasons are:

1. As warm air moves toward a cold region, it tends to be cooled; cold air moving toward warm regions tends to be warmed. When cold air moves across a warm sea surface, this retardation may be as great as 50 percent; and in the special instance of Arctic or polar air from Labrador moving southeast over the warm ocean currents, this retardation may exceed 50 percent. Both of these types of advection retard the movement of isotherms, but in the case of cold air moving over a warm surface, the resultant instability which develops produces an even greater effect on the retardation at upper levels.
2. Cold air spreading southward over warmer regions tends to subside, which results in warming; warm air moving over a cold region tends to be lifted, which results in cooling. This latter factor is the more important one, especially at the intermediate levels aloft. Both factors work against the advection of isotherms with the speed of the wind.

Frontal Analysis

The principal reason for the importance of temperature analysis, especially in the vicinity of fronts, is that the vertical variation of the geostrophic wind at any point in the atmosphere is determined by the temperature field at that point; hence, in the free atmosphere the vertical shear of actual wind is largely controlled by the temperature field. Thus, in areas where no temperature data are available, temperature analysis can still be carried out, provided the variation of wind with height is known. For the Northern Hemisphere, the basic relations can be summarized as follows:

1. If the wind increases (decreases) in speed with increasing height, but does not change its direction, contours and isotherms are parallel, with cold air to the left (right) facing downstream.
2. If neither the wind speed nor direction change with height the air is thermally homogeneous.
3. If the wind veers (backs) with increasing height, the isotherms cross the contours in such a way that advection of warmer (colder) air takes place.

Temperature is generally regarded as representative in the free atmosphere where it is relatively unaffected by many local factors which affect surface temperature. This makes temperature analysis at 850 and 700 mb an important adjunct to surface analysis as an aid in determining the representativeness of the surface temperature and in determining the location and vertical extent of fronts.

The frontal patterns expected aloft are discussed in chapter 5 of this training manual. In the case of wave cyclones, the frontal pattern aloft is roughly parallel to that of the surface wave and displaced from it in the direction of the colder air. In the absence of actual data on the upper air chart, this displacement should be consistent with the height of the chart and the slope of the front.

Frontal analysis is usually not carried above the 700-mb level. On charts which do not carry frontal analysis, isotherms are continuous lines. On charts on which frontal analysis is depicted, some of the isotherms in the vicinity of the front may be discontinuous. Isotherms are generally parallel to the front with the tightest packing in the cold air behind the front. (See fig. 7-17.)

Isotherms can assume other patterns in relation to fronts, indicating the relative strength of the front. The weaker the packing or gradient in the colder air, the weaker is the front and the greater is the probability the front has a shallow slope. Isotherms perpendicular to the front indicate little air mass contrast and consequently a weak front.

You should first sketch in fronts using the thermal gradient, wind shift, and tentative surface frontal positions as guides. Of these three, thermal gradient is by far the most important. Surface frontal and upper level frontal analysis must always be consistent and therefore should never be made independent of each other. Data at both levels should be considered in locating fronts at both levels. Raobs plotted on a thermodynamic diagram are one of the best aids in locating frontal positions aloft. (See chapter 5 of this training manual.)

The consistency of current frontal positions with past positions should be checked and then colored in.

APPLICATION OF SATELLITE CLOUD PHOTOGRAPHS

Among the most significant contributions the satellite has made to meteorolgoy has been its application in the identification, positioning, and prognosis of frontal systems. Some information pertaining to the utilization of satellite pictures during surface analysis as an aid in the positioning and identification of fronts was presented in chapter 6 of this manual. The same general considerations apply when analyzing upper air charts.

An additional factor to consider when analyzing an upper air chart is the displacement and slope of frontal systems with altitude. Although above 700 mb the fronts are normally indicated as troughs, the displacement of these systems must be considered at all levels.

Earlier in the chapter we mentioned certain characteristics of frontal cloud bands which provide information helpful in the positioning of upper level troughs. (See fig. 7-18.) Examination of numerous satellite pictures has revealed that the appearance of a frontal band changes abruptly where it is intersected by the 500-mb trough line. In figure 7-18 the dashed line

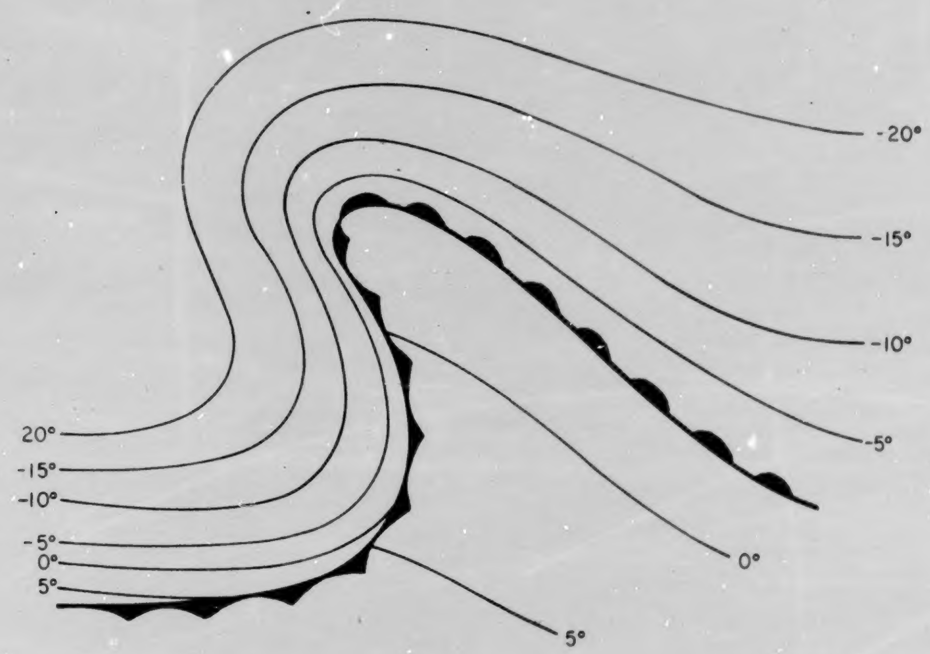

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Figure 7-17.—Isotherm packing at a warm and cold front aloft.

indicates the position of the upper level trough line.

The frontal clouds east of the trough appear much brighter and the band is broader due to upward motion in the southwesterly flow aloft. To the west of the trough line in the region of subsiding air, the clouds tend to dissipate, the front appears more ragged, and there are several cloud-free areas within the frontal band.

Vortices

Cyclonic disturbances produce a variety of spiral cloud patterns, some of which are detached and in some cases disassociated from frontal systems. Fortunately these patterns are not difficult to recognize in satellite pictures.

CUTOFF LOWS.–Vortices associated with upper-level cutoff lows display a variety of patterns, but in the fully-developed state normally cover an area about 10-15 degrees of latitude in diameter. The vortex usually consists of several narrow cloud lines rather than the single band normally seen in extratropical cyclones. (See fig. 7-5.) Since the IR mode represents cloud top temperature rather than brightness, it shows more cloud height detail than the AVCS or daylight mode. (See fig. 7-4.)

The formation and development of cutoff lows is often strikingly apparent from the satellite data in two general ways:

1. When a frontal band separates in advance of a moving 500-mb trough, the equatorward portion of the upper-level trough has cut off, and a closed 500-mb low is forming along the upstream edge of the isolated cloud mass.
2. When lines of cumulus congestus cells (open) are observed to form cyclonically curved lines downstream from an approaching frontal band, an upper-level cyclone is usually forming ahead of the front as illustrated in figure 7-19.

In either of the preceding cases, the location of the upper-level center may be determined by examining the overall cloud pattern in the IR

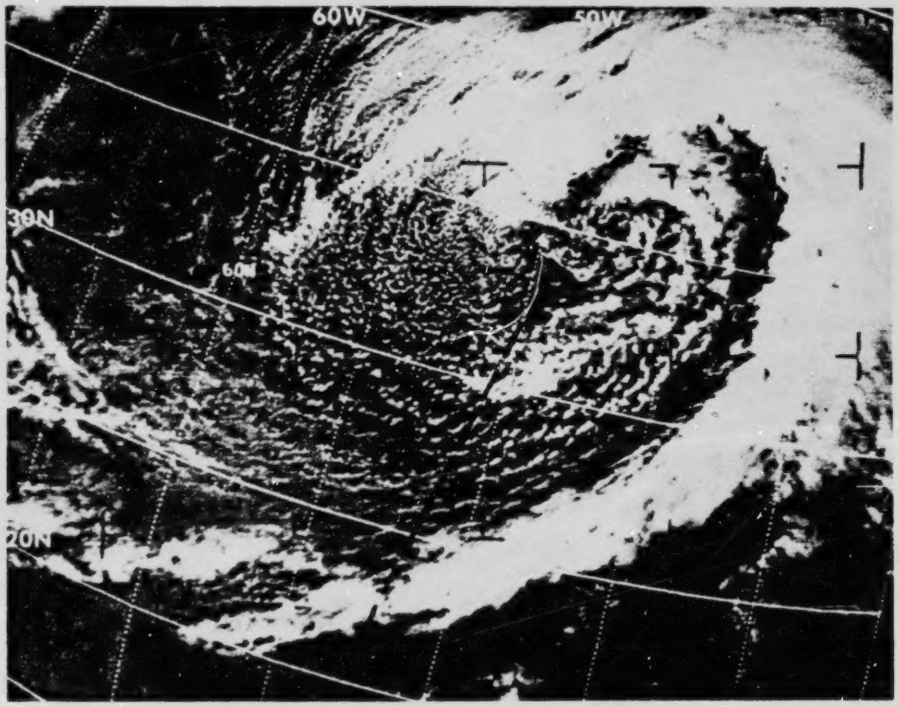

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Figure 7-18.—Change in frontal clouds where an upper level trough intersects a front.

picture. Unless a distinct center appears in low-level cloudiness, the forecaster cannot determine whether or not a closed circulation exists at the surface.

LOW-LEVEL CLOUD PATTERNS.–Low-level cloud patterns frequently encountered in the IR are low-level vortices and fields of stratus or stratocumulus.

The centers of low-level cloud vortices are difficult to locate in the IR picture, due to the fact that there is little temperature contrast between low cloud tops and the earth ocean surface. (See fig. 6-15.) Centers may be especially hard to identify with the older APT receivers, or in dissipating cloud systems. In most cases, it may be possible only to approximate the geometric center of the entire cloud pattern. The centers of weak low-level cyclones may be located with more confidence in the visual data than in the IR.

Cells (closed) appear as rather flat gray; cells (open) appear as a field of mottled gray which has white elements if cumulus congestus clouds are present. It is often difficult to distinguish between cells (closed) and stratus due to the coarse resolution of the IR and small temperature contrast with the surface. The overall pattern and geographic location may be used as guides. (See fig. 6-12.)

APPLICATION OF COMPUTER PRODUCTS

When utilizing computer products for frontal analysis on upper air charts, the same procedures presented earlier in this chapter, and in chapter 6 of this manual, should be followed. Among the more useful computer charts for determining frontal position and intensity is the GG THETA chart.

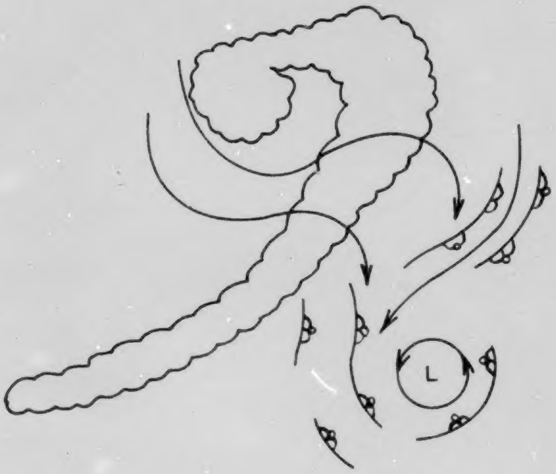

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Figure 7-19.—Formation of 500-mb cutoff low ahead of a front.

GG THETA Chart

The GG THETA chart is a frontal analysis and prognosis arrived at by mathematically identifying and analyzing the direction and magnitude of discontinuities in gradient within the temperature field. The chart is a hemispheric product involving calculated thicknesses between the 700 and 1,000 mb levels converted to mean potential temperature. (See fig. 7-6.)

Use of the GG THETA Chart

When using the GG THETA it must be remembered that the lines on the chart represent the rate of change of potential temperature gradient. The packing of lines (increase in rate of change) may not always be the result of a front. Some of these exceptions will be presented in later paragraphs. One means of checking the existence of a front in the area is to overlay the GG THETA chart with the surface chart. If, however, a front is determined to exist, it can be reasoned that the more lines you have in a particular area, the stronger the contrast will be between the two air masses, hence, the more intense the front will be. For example, if two consecutive GG THETA charts show an increase in the number of lines in a given frontal area, this would indicate intensification of the front.

GG patterns are very useful in determining the location of surface fronts in regions of reasonably dense upper air observations. Areas of frontogenesis or frontolysis will generally be depicted by the GG THETA well before any indication in the synoptic data.

The following paragraphs present a number of conditions under which difficulties might arise and extreme care should be exercised in the identification of fronts on the GG THETA chart.

1. Occlusions may not be depicted due to the lack of strong horizontal thermal gradient in the lower levels, and because the thermal gradient is not perpendicular to the occluded front.
2. If a polar high is shallow and only extends to around 850 mbs the front preceding it may not be evident on the GG THETA because this chart is derived from the 1,000 to 700 mb layer.
3. The GG THETA may present difficulties in identification of frontal zones or the determination of frontal intensification when two fronts are in close proximity, as when one front approaches another. In some cases one front may be absorbed by the other. Since all zones indicated on the GG THETA over land areas are not necessarily fronts, distinguishing of the true front may be hindered by the addition of a zone caused by topography, land vs water, etc.

In the winter, over Asia in particular, where cold pockets of air are trapped in the lower levels, it will be indicated on the GG THETA as an area of sharp thermal contrast. Of course the indicated frontal zone is in actuality the mountain barriers that have pocketed the cold air. Also in the summer, the thermal low found over Mexico will indicate a frontal zone on the GG THETA but close inspection by the analyst or forecaster will normally reveal no frontal type weather.

4. The use of the GG THETA over extensive water areas should be carefully evaluated. Just as with other types of charts, the number of surface and upper air reports available, their accuracy, and whether the reports were representative must be considered. If a front exists,

there should be packing of the isotherms indicated on the 850-mb chart. In addition, the 500- to 1,000-mb thickness analysis should show a concentration of thickness lines in areas of possible frontal location.

AUGMENTING DATA

Radiosonde reports may be used to further improve a frontal analysis. As mentioned previously, the computer will not draw in fronts. Also the computer will not kink isobars or contours unless adjacent reports are near enough together to indicate a kink. Observing wind shift and temperature change on radiosonde reports may all pinpoint accuracy on frontal location.

Use of the International Analysis Code (IAC) will make it possible for comparison of your analysis with the analysis of another individual, thereby allowing you to take the better parts of both analyses.

BLOCKS

Experience has shown that about one-third of the time the distortion of the westerlies is so great that the rules covered in chapter 4 on zonal and meridional flow patterns are totally inapplicable. A condition that exists at intervals is known as the blocking situation (also called the cutoff high). Types of blocks are discussed in chapter 4 of this training manual. It is rather difficult for the analyst to determine a blocking condition from a single chart. The method of detecting existing blocks depends primarily on the use of 2- to 3-day running means of wind and pressure aloft, particularly at the 500-mb level.

Satellite Depiction of Upper Level Blocks

A good example of the use of satellite photographs to orient the position of a block is illustrated in figure 7-20.

A large ridge in the form of an omega block often has minor ridges within the large scale flow. Line AB, in both figures 7-20 and 7-21 illustrates this feature. The major ridge line shown in both figures runs from position C to D. It can be seen that the minor ridge line closely approximates the northern edge of the cloud band, spiralling into the vortex at position E in figure 7-20. Cloud patterns in satellite photographs correlate with the 1,000- to 500-mb thickness patterns. This is because the cloud patterns are basically the result of integrated motions throughout the atmosphere and not at any particular level.

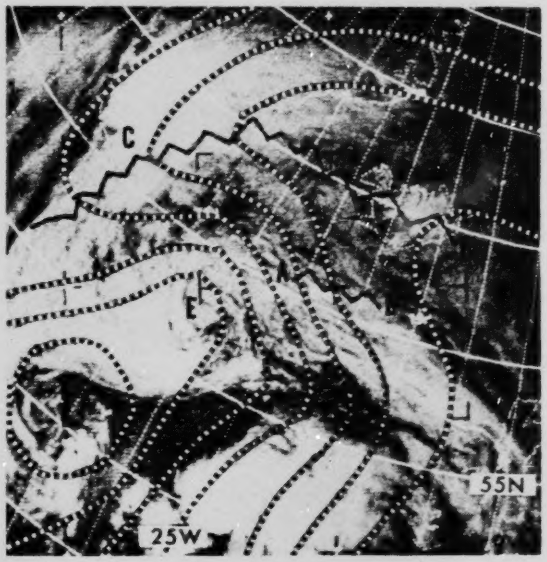

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Figure 7-20.—500-mb analysis and satellite photograph showing an omega type blocking high.

Use of Computer Products

The thermal characteristics related to blocking situations were described in chapter 4 of this manual. By studying the computer 500-mb analysis (figure 7-9), the 1,000 to 500 mb thickness charts, the vertical velocity analysis, and other tropospheric and stratospheric analyses, considerable assistance may be obtained in determining whether a blocking situation exists. A complete discussion of blocks and blocking situations may be found in the Navy publication Handbook of Single Station Analysis and Forecasting, NA 50-1P-529.

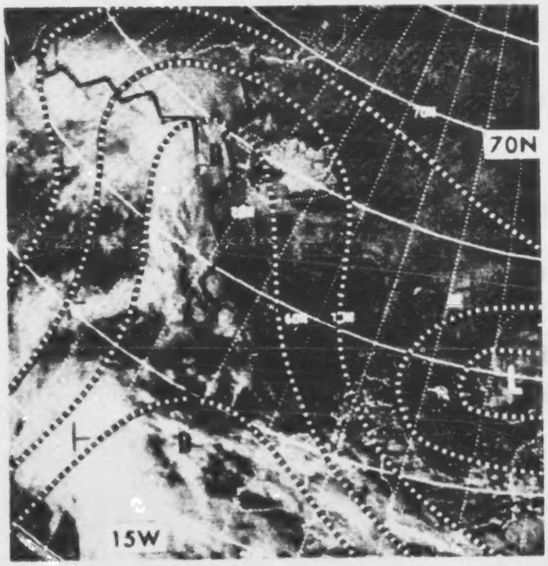

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Figure 7-21.–500-mb analysis and satellite photograph showing an omega type blocking high.

The author suggests in this publication that for single station forecasters it has been found adequate to define blocking solely in terms of closed anticyclonic patterns which persist at the 500-mb level for 2 days or more north of a certain critical latitude. In the Atlantic Ocean this was found to be 40° N; in the Pacific Ocean 35° N. In both cases, when a closed anticyclone appeared on the 500-mb chart north of the critical latitude and persisted for 2 days or more it was attended by the familiar features of blocking action.

The blocking types, as developed empirically for both the Atlantic and Pacific Oceans, have a number of common features which permit a sketch of a typical blocking pattern. These features are shown graphically in figure 7-22.

Blocks should be looked for during their most frequent seasons of occurrence in late winter and early spring.

MOISTURE ANALYSIS

Lines of constant mixing ratio or isodrosotherms may be drawn on 850- and 700-mb charts for the purpose of delineating the moist and dry tongues in meridional flow patterns. The extent and intensity of moisture patterns are of particular importance in areas where the atmospheric processes producing clouds and/or precipitation predominate.

Isodrosotherms are discontinuous at fronts. Moist tongues are most pronounced in warm maritime air masses, while dry tongues are most marked in cold continental air masses. For obvious reasons, an isodrosotherm cannot cross the isotherm with the same temperature label.

The National Meteorological Center (NMC) provides a Composite Moisture Index Chart. This is a four panel chart for the conterminous United States. The four panels present information pertaining to index conditions, precipitable water, freezing level, and relative humidity. Details on the construction of this computer produced chart and transmission times may be found in the National Weather Service Forecaster's Handbook, WBFH No. 1. Local weather units may utilize this computer produced data as an aid in determining areas close to saturation.

On a working chart light green is the color customarily used for isodrosotherms, with light green shading for moist tongues, becoming progressively darker toward the axis of the tongue.

The troughs and ridges in the isotherm patterns are called cold and warm tongues, respectively, and in the moisture pattern they are referred to as dry and moist tongues, respectively.

INDEXES

The types and characteristics of zonal indexes are described in chapter 4 of this training manual. A method of computing zonal indexes is also covered in that chapter. Indexes are best identified on mean charts. However, there are indications on the present upper air charts (700 mb and above) which present identifiable features of the indexes. A brief review of the indexes and some methods for identifying them are presented here.

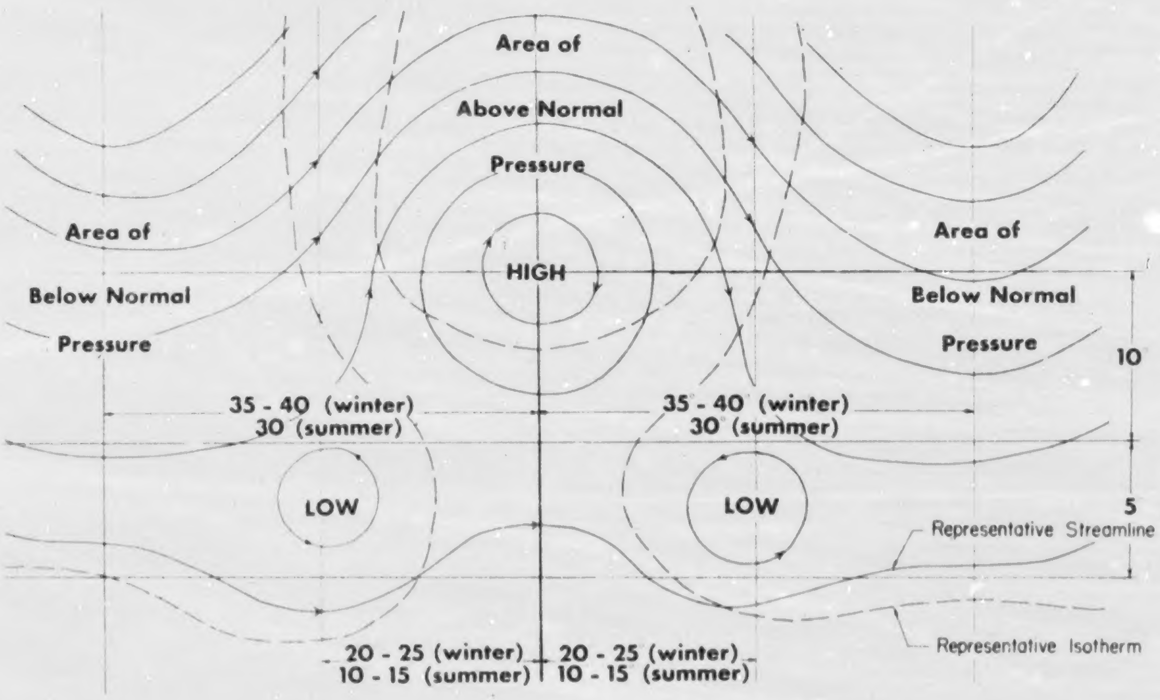
Area of
Above Normal
Pressure
HIGH
Area of
Below Normal
Pressure
Area of
Below Normal
Pressure
10
35 - 40 (winter)
30 (summer)
35 - 40 (winter)
30 (summer)
LOW
LOW
5
Representative Streamline
Representative Isotherm
20 - 25 (winter)
10 - 15 (summer)
20 - 25 (winter)
10 - 15 (summer)

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Figure 7-22.—Characteristic 500-mb surface streamlines and dimensions for oceanic blocking (all measures in degrees of longitude and latitude).

Characteristics

In determining the index pattern, we are primarily concerned about the fluctuations of the principal band of westerlies between 35 and 55 degrees north latitude. From computations as described in chapter 4 it is found that zonal indexes may be computed from the indicated strength of these westerlies.

A high index pattern, which is the normal pattern, connotes a westerly flow with high pressures to the south and low pressure to the north. A primary characteristic of this pattern is the more or less regular arrangement of contours aloft in a west to east direction and the presence of quasi-stationary flat troughs of long wavelength through which moves rapidly a series of fast, shallow lesser troughs which are warmer than their environment and which have little vertical development. A group of these wave perturbations superimposed on the westerly flow correspond to a family of wave cyclones at the surface. There is little northward or southward movement of pressure systems nor rapid development of centers. In short, there is no interchange of contrasting air masses from north to south. Therefore, with a high index circulation, surface lows move rapidly eastward without developing into major systems.

A high index pattern is the normal condition as may be evidenced by the west-east orientation of the 700-mb contours on mean charts. An additional characteristic is the occurrence of pressure near normal with only small deviations.

A low index pattern consists of large north-south components of flow with a general decrease or even reversal of the pressure gradient of low pressure to the north and high to the south as associated with the high index.

The primary characteristic of this pattern is large amplitude perturbations in the westerlies and the existence of meridional flow over large areas, which means that widely contrasting air masses are interchanged. In some cases a closed low circulation may exist aloft in the mid-latitudes. Troughs are slow, deep, and colder than their environment and have marked vertical development. Under these conditions surface lows will move less rapidly, unfavorable weather will spread over large areas, and consequently clearing will occur less rapidly than when a high index circulation is present. Similarly, aloft, the trough will have a smaller west to east component of movement.

Index Type from Radiosonde Analysis

The use of radiosonde analysis in locating fronts was discussed in chapter 5 of this training manual. The radiosonde can also be a valuable aid in identifying the index situation through the variation of temperature and height of the tropopause in middle latitudes.

The variation can be related to surface air masses. At approximately 45 degrees north latitude the northward influx of tropical air is associated with a high cold tropopause of the order of 15 to 18 km in height and -70° to 80° C temperature. A low warm tropopause of 7 to 9 km and temperatures -50° to -55° C is associated with a southward surge of Arctic air and an intermediate tropopause of 11 to 13 km and -60° to -70° C accompanies the advent of maritime air which is the normal case. The greater the deviation of the tropopause from the mean position, the more intense is the northward or southward push of air.

Since the presence of tropical or polar air in the middle latitudes is associated with a low index type, it follows that the tropopause position is a good key to the circulation index. The appearance of a distinct change in lapse rate at the tropopause level is linked with a low index pattern, and a gradual or rather indistinct change in the lapse rate with the tropopause level above 12 km indicates the existence of high index conditions.

COORDINATION OF UPPER AIR DATA AND SURFACE MAP

Certain implications from the above data and other data obtained from upper air observations can be employed to determine the configuration of the current map. Primarily, the overall picture of the current situation is obtained by visualizing the index pattern. Conclusion as to whether high, low, or transitional index should be deduced from consideration of the factors stated in chapter 4, consideration of the above, and from the following features. Either 700- or 500-mb levels may be used.

1. Deviation of the 700-mb heights and temperatures from normal and of the tropopause height and temperature.
2. Past weather.
3. Direction changes in the flow at 700 mb.
4. The height tendency at 700 mb or some higher level.
5. Types of air masses present.

For instance, a continued westerly flow with 700-mb heights and temperatures near the normal and no large change would allow an assumption of a high index. On the other hand, large deviations from the 700-mb normal height and temperatures and large height change values would indicate other than a high index pattern.

COORDINATION OF SATELLITE AND COMPUTER PRODUCTS

Satellite photographs and computer products such as those described earlier in this manual provide a valuable source of assistance in determining index conditions.

Satellite Photographs

The existence of low index situations, commonly associated with baroclinic conditions, are evident on satellite photographs by the existence of frontal systems, vortices, and other forms of convergent weather. High index conditions will be characterized on the photographs by a predominance of fair weather conditions.

Computer Products

There are a variety of computer products available, primarily in the form of upper level charts, which will provide the analyst assistance in determining index conditions. The use of these charts has been presented previously in this manual. It would prove redundant to attempt to list the various charts since they change from time to time and are listed in the Computer Products Manual.

LONG WAVES, TROUGHS, AND RIDGES

Two superimposed wave patterns can generally be distinguished on upper air charts (700 mb and above). Slow moving features designated as long wave troughs and long wave ridges lie under a system of fast moving disturbances designated as short wave troughs and short wave ridges. The identifying features are discussed in chapter 4 of this training manual.

Long wave patterns are best detected and identified on mean or other type charts. All wave patterns are classified according to their wavelengths as either long or short waves. Long waves vary in length from about 60 to 120 degrees longitude and their normal movement is from west to east about 5 degrees longitude per day, but long waves can be stationary or even retrogress to the west.

Great care in tracking trough and ridge lines on successive maps is necessary if proper continuity is to be maintained. This is especially true of short waves, which appear as perturbations of small dimension moving rapidly along the long wave pattern at a speed of 10 to 20 degrees per day. Because of the sparseness of upper air data in many areas these short waves are easily overlooked or carelessly smoothed out of the analysis. In isolated areas, the principal clue to the passage of a short wave trough is a slight backing of the wind for a brief time. The long wave trough intensifies when overtaken by the short wave trough and this intensification often results in surface cyclogenesis. The best synoptic clue as to the existence and location of short waves in denser networks is the 12-hour height change pattern. A small closed pressure change center is often associated with a short wave trough or ridge.

Figure 7-23 illustrates common occurrences of short waves, long waves, and a surface low. Note in this figure that the slope of the surface pressure system varies from near zero in the Mediterranean, where the presence of cold air masses is extremely rare, to 8 to 10 degrees longitude off the eastern coasts of the United States and Asia.

Typical wave patterns of isobars and isotherms of various levels in the zone of the westerlies is shown in figure 7-24. The most significant features of this observed pattern are that the surface isobars and isotherms are most frequently about 135 degrees out of phase. Using either the hydrostatic equation or the thermal wind equation, it is seen that this surface distribution produces a vertical tilt to the trough and ridge lines and a decrease in the amplitude of both the isohypses and the isotherms until both are in phase. This normally occurs at about the height of the 600-mb surface, which is also the average height of the level of nondivergence. About the level at which the isohypses and the isotherms are in phase, the amplitude of the horizontal isohypses and isotherms will increase upward to the tropopause. Troughs and ridges normally slope very little above 500 mb. From the distribution of the wind speed profile, it is seen that below the 600-mb surface, waves will normally travel faster than the zonal wind, and somewhat above that level, the zonal wind is usually higher than the wave speed. The corresponding fields of convergence, divergence, and vertical motions are shown for reference.

Satellite and Computer Products

Satellite photographs and computer products provide an excellent means of determining the present as well as future positions of long and short wave troughs. The application of these important devices to analysis as presented earlier in this manual should not be overlooked.

VORTICITY ANALYSIS

The determination of relative vorticity is covered in detail in chapter 4 of this training

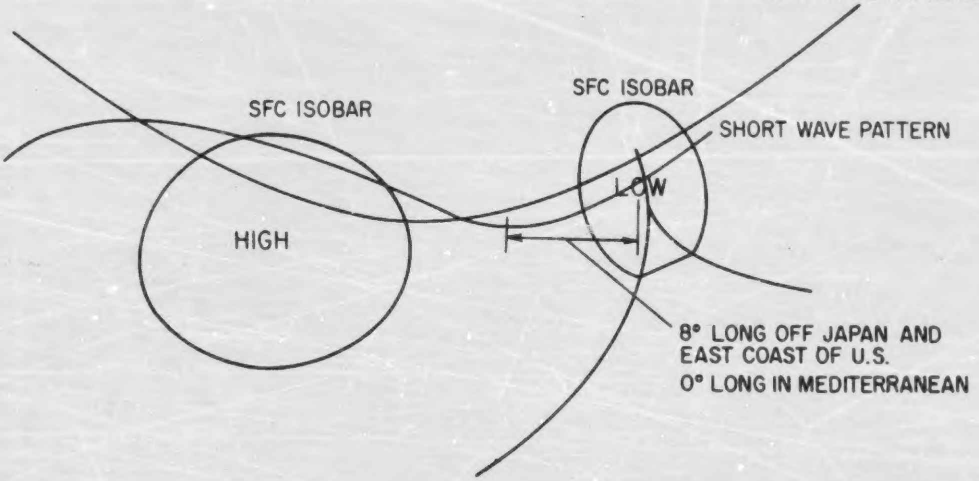
LONG WAVE PATTERN
SFC ISOBAR
SFC ISOBAR
SHORT WAVE PATTERN
LOW
HIGH
8° LONG OFF JAPAN AND
EAST COAST OF U.S.
0° LONG IN MEDITERRANEAN

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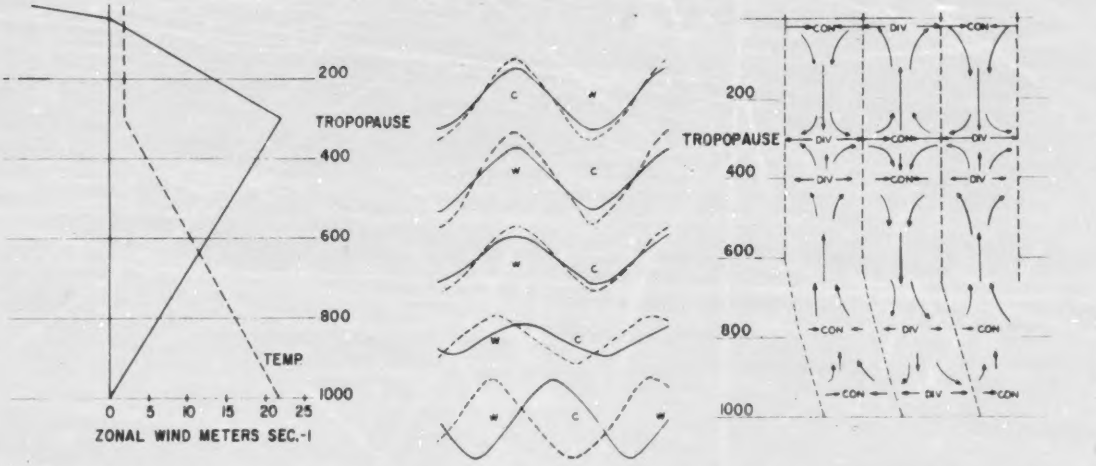
Figure 7-23.—Short waves, long waves, and surface features.


200
TROPOPAUSE
400
600
800
1000
TEMP
0
5
10
15
20
25
ZONAL WIND METERS SEC.-1
C
W
W
C
W
C
W
C
W
C
W
200
TROPOPAUSE
400
600
800
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TYPICAL VERTICAL WIND SPEED AND TEMPERATURE PROFILE

PLAN VIEW AT VARIOUS LEVELS OF THE ISOTHERM-STREAMLINE PATTERN

VERTICAL PROFILE ALONG THE WESTERLIES SHOWING THE VERTICAL WIND AND DIVERGENCE PATTERN

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Figure 7-24.—Typical wave patterns in the westerlies.

manual. Vorticity charts for present and prognostic conditions are currently being transmitted at regular intervals on the National Facsimile Network.

In practice, actual vorticity values are rather difficult and time consuming to compute. With the advent of computers, this computation is performed with speed and relative accuracy. In practical analysis you should remember that positive vorticity is a normal occurrence in advance of troughs in the middle and upper troposphere. A good indication of cyclogenetic development at sea level can be deduced when an upper trough with positive cyclonic vorticity in front of it overtakes a frontal system in the lower troposphere. You should also remember that anticyclonic vorticity is negative and the areas from the rear of the troughs to the next ridge lines are favorable for anticyclonic development.

Satellite and Computer Products

SATELLITE PHOTOGRAPHS.–An important means of identifying areas of positive and negative vorticity is the satellite photograph. Comma shaped areas of clouds described as Positive Absolute Vorticity (PVA) Maxima and vorticity centers occur in the cold air to the rear of, or around the periphery of, cloud vortices related to extra tropical systems. Figure 7-25(A) shows a DRIR readout with a vortex at position A and a frontal band extending from position B to C. Behind the frontal band, a PVA MAX is shown at position D rotating around the vortex in the cold air.

As illustrated in figure 7-25(B) thickness lines are generally normal to the cloud band of the comma. It is important to identify these features since if a PVA max overtakes an active baroclinic zone, cyclogenesis is likely to occur.

COMPUTER PRODUCTS.–A number of different computer charts (SD 500 ANAL, 500 MB HT ANAL, SL 500 ANAL, H5-10 ANAL, etc.) are available for use in determining vorticity conditions. Among these is the "Vorticity Advection Chart." Through the use of this chart, a forecaster can study the progression of the positive vorticity centers and carefully examine the relationship of these centers in conjunction with the long wave troughs located on the SL 500 MB chart, and the surface chart. The effects of vorticity on weather were presented in chapter 4 of this manual.

The tracking of positive vorticity centers has become a very significant aid to the forecaster when trying to determine the intensity changes of long wave troughs and surface lows. Through the use of computer charts along with other available forecasting aids this task will become somewhat less difficult. The utilization of vorticity in forecasting will be presented in the pertinent forecasting chapters of this training manual.

ISOTACH AND JETSTREAM ANALYSIS

Where wind data are sufficiently concentrated, isotachs are drawn to locate the axis along which the wind speeds are greatest. This axis is the ridge line connecting the various centers of relative maxima in the isotach pattern. The isotachs are almost parallel to the isohypses and the closed centers are usually elongated ellipses with the long axis parallel to the isohypses. Two idealized contour and isotach patterns are shown in figure 7-26. The one on the left shows a speed maximum at the long wave trough line, the other shows maxima at the ridge lines. These centers normally move from west to east with a speed less than that of the winds themselves, but greater than that of the long waves.

The isotach ridges on any particular constant pressure chart represent the intersection of that surface with a meandering current of fast moving air, known as the jetstream. Since wind speed varies in the vertical as well as the horizontal, the core of the jetstream fluctuates vertically as well as horizontally from one meridian to another. It usually lies between the 300- and 200-mb levels, nearer to the 200-mb level in lower latitudes and nearer the 300-mb level in higher latitudes. Its width varies from 250 to 400 miles, being narrower during periods of low index and wider during periods of high index. Chapter 4 of this training manual contains illustrations of the jetstream in relation to the polar front and the tropopause.

The 300- and 200-mb charts give the best representation of the jetstream, but it is often

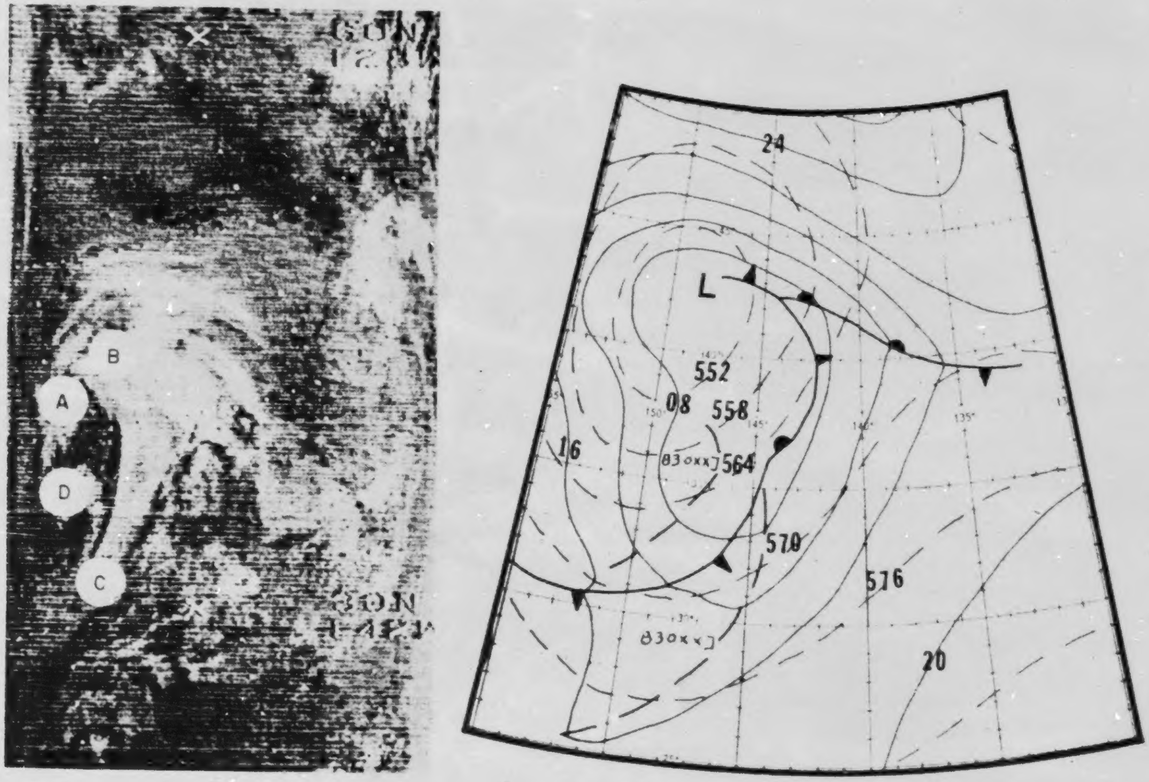

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Figure 7-25.–(A) DRIR readout showing a vortex, frontal band, and PVA Max; (B) corresponding surface chart with superimposed thickness lines.

well defined down to 500 mb. Below this level the strong meridional temperature, particularly those associated with the polar front in the winter, causes the intensity of the speed maximum to decrease rapidly in the lower level of analysis.

Above the jet core, the vertical axis of the jetstream tilts equatorward. From 500 to 300 mb, it has almost no tilt, and is usually located directly beneath the line of maximum horizontal temperature gradient at 200 mb; hence, the 500-mb isotach analysis and the 200-mb temperature analysis are both very important aids in locating the 300-mb jet below the jet core. The lack of tilt in the vertical axis of the jet implies, by virtue of the thermal wind relation, that maximum isotherm concentration coincides with the jet axis. At 500 mb, the isotherm concentration is often associated with the polar front. The jet axis at this level seems to be around the -20°C isotherm on the warm side of the zone of maximum temperature gradient.

In the horizontal, the jet axis also follows quite closely the contours. At 500 mb it lies near plus or minus 60 meters of the 5,610-meter contour. Project AROWA found in one study that the jet maximum occurs 82 percent of the time between the 8,960- and 9,240-meter contours at 300 mb.

Two exceptions to the foregoing should be noted. At low wind speeds, contours and isotachs may cross at large angles, and where

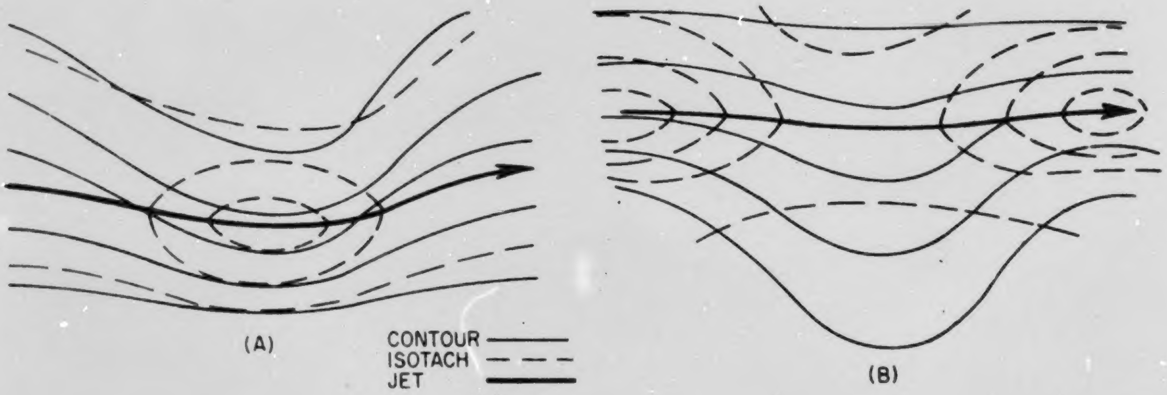

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Figure 7-26.—Common contour and isotach patterns. (A) Speed maximum at long wave trough line; (B) speed maximum at ridge line.

contours converge (upstream from a speed maximum), the jet axis and isotachs tend to cross from high to low, the reverse being the case downstream where the contours diverge. Deepening and cyclogenesis occur downstream and to the left of the jet maximum at a trough and upstream and to the right of the maximum at ridges. The relation of contours to the jet axis also accounts to some extent for the appearance of the meridional and easterly jets and the forked jets associated with cutoff lows and blocking highs.

LIFE CYCLE AND SEASONAL POSITIONS OF THE JETSTREAM

The period of life cycle of a particular jet determines the characteristic structure of the jet at a specific time over a given region. Although it may be convenient to visualize the jetstream as a tube of high wind speeds meandering around the hemisphere, it is clear that its character generally varies from region to region.

The seasonal aspects of the jetstream were discussed in chapter 4 of this training manual. In summary, it may be said that the strength of the jetstreams is greater in winter than in summer. The mean position of the stream shifts south in winter and north in summer with the seasonal migration of the polar front. As the jetstream moves south, its core rises to a higher altitude and, on the average, its speed increases. In the winter the jetstreams are often found as far south as 20 degrees N. The core of the strongest winds in the jetstream is generally found between 7,620 and 12,190 meters, depending upon the latitude and the season.

APPLICATION OF SATELLITE CLOUD PHOTOGRAPHS

The location of the jet stream or maximum wind zone can be accurately positioned by examining the cloud patterns shown in satellite pictures. Vertical and horizontal motions in the upper troposphere in the vicinity of jet streams have a very marked effect on the distribution of cirrus clouds in their vicinity. (See fig. 7-27.)

Because of this, it is possible to determine the location of such wind maxima from cloud pictures taken by weather satellites. Cirrus clouds predominate on the equatorial side of the jet stream and in the anticyclonically-turning portion of the jet. The poleward boundary of the cirrus is often very abrupt; it lies under or slightly equatorward from the jet axis, and frequently casts a shadow on the lower clouds which is clearly visible in satellite pictures.

Over the oceans, where the jet stream curves cyclonically, the differences in stability on each

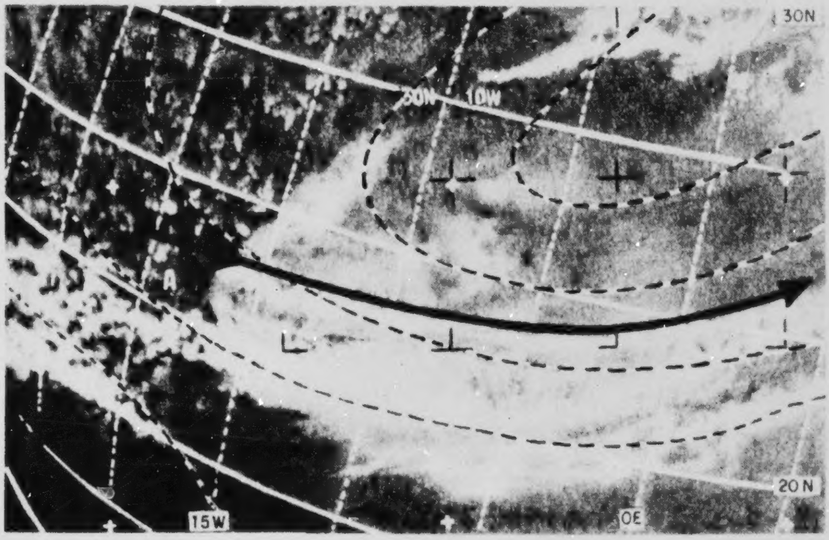

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Figure 7-27.—200-mb analysis showing transverse lines in jet stream cirrus over Africa where lower clouds are absent. The southern edge of the lines trails off to the west due to slower wind speeds farther away from the jet core.

side of the core are reflected in the appearance of the clouds. On the left side, looking downstream, cold temperatures and unstable air occur resulting in great vertical development of the convective clouds in the open cellular patterns. (See fig. 7-28.)

One of the major problems facing the analyst is differentiating between jet stream cirrus and instability lines of Cb activity. In both cases, the display may be white, representing emission from high, cold cloud tops. The forecaster should examine the equatorward edge of the cloud shield to determine whether the boundary is sharp or diffuse.

The main jet-stream cloud features are long shadow lines, large cirrus shields with sharp boundaries, long cirrus bands, cirrus streaks, and transverse bands within cirrus cloud formations. Once a jet stream has been identified on a cloud photograph, it is possible to deduce other meteorological information, such as wind direction, wind shear, direction of horizontal temperature gradient, and areas where clear air turbulence is possible. The Guide for Observing the Environment with Satellite Infrared Imagery, NWRF F-0970-158, and Application of Meteorological Satellite Data in Analysis and Forecasting, Technical Report 212, provide an excellent source of additional information on locating the jet stream through the use of satellite photographs.

APPLICATION OF COMPUTER PRODUCTS

The most useful computer charts available for analyzing the jet stream are the 200 mb, 300 mb, isotach, and wind shear analysis. The isotach analysis is especially useful for determining positions of jet maxima. The characteristics of the jet stream were presented in chapter 4 of this manual. By utilizing this descriptive data along with the available computer charts accurate positioning of the jet stream will be practically assured. Isotach analysis will be described in greater detail in the following paragraphs.

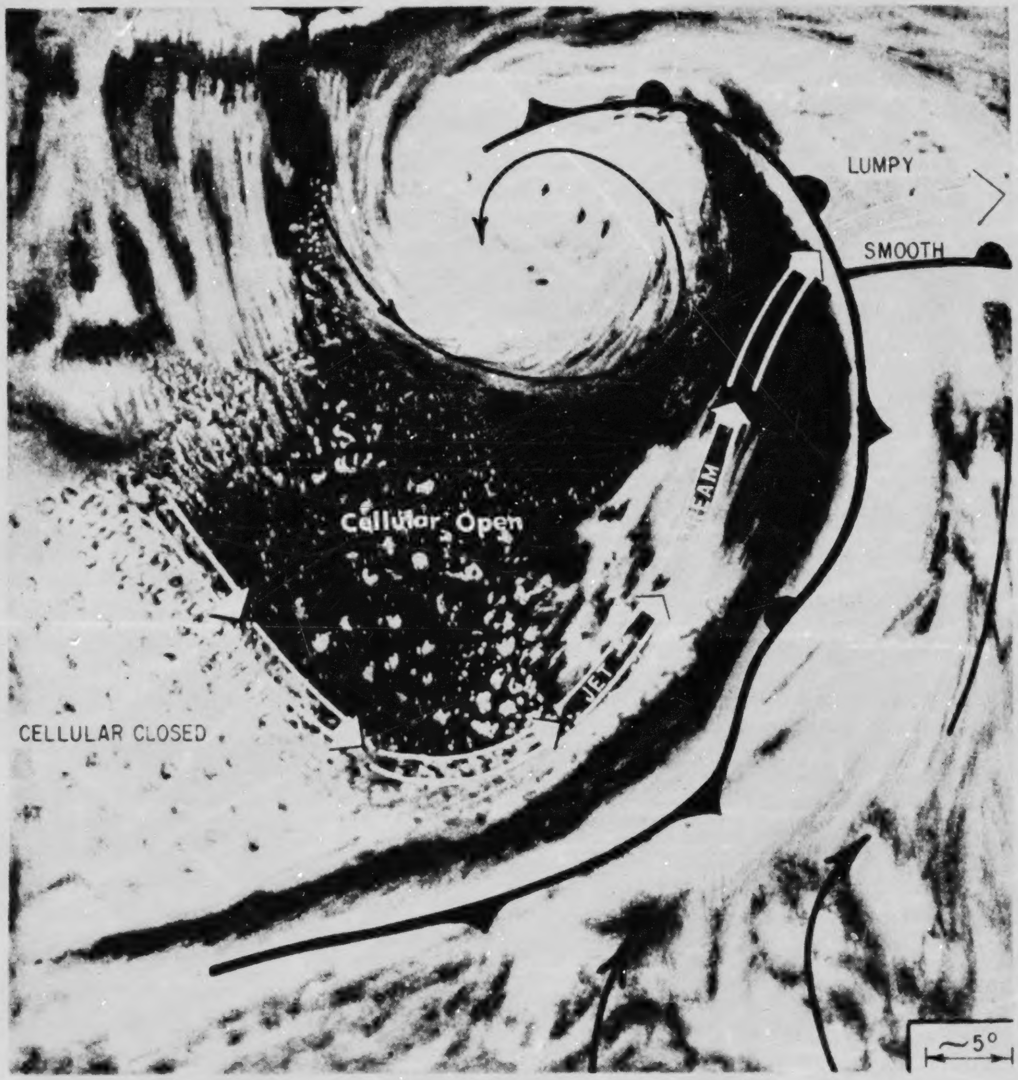
LUMPY
SMOOTH
Cellular Open
JET STREAM
CELLULAR CLOSED
~5°

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Figure 7-28.—An idealized illustration showing a jet stream over the dividing line between open and closed cellular clouds. The jet stream east of the upper air trough is shown crossing over an occluded front.

ISOTACH ANALYSIS

The first step in isotach analysis is the computation of wind speeds in areas of sparse or little wind data. This can be accomplished by the use of gradient and geostrophic wind scales. Many instances have occurred in which observed winds do not agree with geostrophic or gradient computations, thereby placing this method of augmenting wind speed data into the same category as the extrapolation of data upward from a lower level–it is better than no data at all. The following items should govern the computation of wind speeds:

1. Use geostrophic or gradient wind scales constructed for the map projection, if available.
2. First select those points where the wind is most likely to be geostrophic, that is, where contours are straight and parallel or with little curvature. Then measure the geostrophic wind.
3. Next, choose areas where contours are curved but approximately parallel and measure contour curvature from overlay, then estimate the correction for trajectory curvature, and compute the gradient wind from available scales.

In the cases of number 3, a recent statistical study has shown that only in cases of cyclonic curvature do you need to correct the geostrophic wind to the gradient wind. In anticyclonic cases, the geostrophic wind seems to give as good an approximation to the true wind as does the gradient wind.

Once the wind speed data have been augmented as outlined above, isotach analysis should proceed as follows: (The recommended interval for isotachs is 20 knots.)

1. Begin the analysis in an area with dense reports and draw in the 60-knot isotach. If you are using either the 300- or 200-mb chart, use the 500-mb chart as an aid; it has greater coverage.

2. Draw in the remaining isotachs at 20-knot intervals and identify the centers of speed maximum from previous charts (continuity).

3. Sketch in the jet axis. Interpolate the jet axis in areas between centers, following the contour and isotherm pattern.

4. Check for consistency with jet axis and isotherm patterns above and/or below the level being analyzed.

After completion of the isotach analysis, the isotach patterns should be pictorial. One recommended procedure is as follows:

1. At first, draw the principal jet axis as a heavy purple line with arrow indicating the flow direction. Indicate secondary jets and jet fingers with dashed purple lines.
2. Shade regions with wind speed less than 20 knots (also marked with an S) purple and regions with wind speeds greater than 80 knots (at 500 mb, 60 knots) green. Intense jet centers may be emphasized by increasing the heavy shading toward the center.

Centers of wind maxima usually propagate downstream at a rate greater than that of the long wave pattern, but less than that of the winds themselves; hence they will have movement relative to the wave pattern and appear to move through it, much like the short wave patterns do. Occasionally centers will remain stationary but have never been known to retrogress upstream.

OTHER AIDS IN LOCATING THE JETSTREAM

The following features are used at the National Meteorological Center to focus attention on areas where the jetstream might be located.

1. The jetstream lies vertically above the maximum temperature gradient in the midtroposphere as indicated for example by the 500-mb temperature field. It will be located just to the south of the greatest thermal gradient.
2. The jetstream reaches its maximum intensity slightly below or at the tropopause. Thermal gradient is reversed above this level.
3. The cyclonic shear to the left of the jet (looking downstream) is stronger than the anticyclonic shear to the right.
4. The jetstream is also a streamline and as such tends to follow the same height contour. From this is follows that the axis of the jet on a constant pressure surface is directed toward

lower heights when wind speeds increase along the jet and vice versa.

5. Information from satellite pictures aids in locating the jetstream as covered in chapter 14.

TROPOPAUSE ANALYSIS

The tropopause is the boundary of transition between the troposphere and the stratosphere. FMH No. 3, Radiosonde Observations, NA 50-1D-3, gives criteria for selection of the tropopause levels at pressure lower than 500 mb from radiosonde data. A template is available, Form FMH 3-31B1, for use in selecting the tropopause level. FMH No. 4, Radiosonde Code, NA 50-1D-4, gives criteria for encoding tropopause data.

Formerly, it was believed that the tropopause was a single unbroken layer extending from the Equator to the poles and sloping downward toward the poles. Data gathered subsequent to and during World War II have served to invalidate this theory. There have been found to be at least three more or less distinct tropopauses, which form leaflike or overlapping structures between which jet cells of maximum wind are found.

The three generally accepted tropopauses are the subtropical tropopause found at about 25 degrees N near 18,290 meters or around 100 mb, midlatitude or temperature tropopause at 35 to 40 degrees N near 12,190 meters (around 200 mb), and the subarctic tropopause near 9,145 meters (around 300 mb). In general the tropopauses are found at greater heights in summer than in winter.

Each of the tropopauses slopes downward toward the north. For a short distance the subtropical tropopause tends to overlap the temperate tropopause and the temperate tropopause tends to overlap the subarctic tropopause. These regions of overlap are characterized by double or complex tropopauses, and the two main regions of overlap or discontinuity are associated with the midlatitude and subtropical jetstreams.

These three tropopauses are characterized not only by height and pressure but also by potential temperature. Within approximately ±10° C the potential temperature in winter is 390° K for the subtropical tropopause, 350° K for the temperature tropopause, and 310° K for the subarctic tropopause. The characteristic potential temperature may be of value in locating the tropopause on an atmospheric sounding when the sounding is characterized by many inversions or by an irregular lapse rate with no inversions.

There are many soundings on which the locating of a generally acceptable tropopause is very easy. These cases include those where a single strong inversion or marked stabilization of the lapse rate occurs. However, there are many other soundings where it is quite difficult to determine where the tropopause ends and the stratosphere begins. In these cases, the lapse rate may just gradually become more stable without any prominent sudden stabilization of the lapse rate with height, or there may be more than one point of stabilization; then it becomes a controversial problem to decide which point should locate the tropopause. The variation in structure in the tropopause region has led to many different ideas on how to define and analyze a tropopause. The conflict between the definitions officially advocated or used in different countries was finally compromised by the WMO who have standardized an arbitrary definition of the tropopause for operational use in internationally exchanged upper-air sounding reports.

The present WMO definition does not attempt to settle the question as to what the tropopause is, which remains controversial. It defines, rather, an objective technique for locating the lowest height in the atmosphere where the lapse rate first decreases to an average lapse rate of 2° C/km for a 2-kilometer layer. While the physical significance of this lowest height remains to be found, it does establish an international reference height to which other atmospheric phenomena can be empirically related. The main advantage is that the definition is objective, in that it allows the same height to be consistently selected by all technicians from a given sounding.

WMO TROPOPAUSE DEFINITION

As noted above, objective criteria for determining the tropopause height(s) transmitted in upper-air observations have been internationally standardized by the WMO. Within its definition, provision is made for identifying two or more tropopauses on a sounding. The WMO definition for the tropopause is as follows:

1. The "first tropopause" is defined as the lowest height at which the lapse rate decreases to 2° C/km or less, provided also that the average lapse rate between this height and all higher altitudes within 2 kilometers does not exceed 2° C/km.

2. If above the first tropopause the average lapse rate between any height and all higher altitudes within a 1-kilometer interval exceeds 3° C/km, then another tropopause is defined by the same criteria as under 1 above. This second tropopause may be either within or above the 1-kilometer layer.

3. There are two qualifying remarks attached to the definition. They are:

a. A height below the 500-mb surface is not designated as a tropopause unless it is the only height satisfying the definition, and the average lapse rate in any higher layer fails to exceed 3° C/km over at least a 1-kilometer layer. Further, the sounding must reach at least the 200-mb pressure surface. (The intention here is to admit that a discontinuity in the lapse rate at a height below the 500-mb surface is a tropopause only if it is reasonably certain that no other choice is possible.)

b. When determining the second or higher tropopauses, the 1-kilometer interval with an average lapse rate of 3° C/km can occur at any height above the conventional tropopause and not only at a height more than 2 kilometers above the first tropopause.

RADIOSONDE ANALYSIS

Figure 7-29 illustrates how to apply the WMO tropopause definition to a sounding. The first tropopause is defined by criterion 1 of the definition. This criterion is satisfied above point B of figure 7-29; that is, the average lapse rate between points B and D, the next 2 higher kilometers above point B, is less than 2° C/km for all points of the 2-kilometer lapse rate. Thus, the first tropopause is established at point B of the sounding.

There is also a possibility of a "second" tropopause at point D. To find out, criterion 2 is used. This criterion requires a 1-kilometer layer with a lapse rate greater than 3° C/km below the second tropopause height. This requirement is met through the layer CD, and in this particular case, qualifying remark 3 (b) applies also. When criterion 2 is met, then criterion 1 is again used to determine the location of a higher tropopause. Above point D, the lapse rate DE again decreases to less than 2° C/km for the next two higher kilometers. Thus, a second tropopause is determined at point D.

Tropopause heights and temperatures are currently being placed on the 35,000-foot upper level wind charts and transmitted over the National Facsimile Network.

COMPUTER PRODUCTS

NMC provides a computer derived tropopause and wind shear analysis over the facsimile network which may prove valuable as an aid in positioning the tropopause. This chart is a two panel presentation for the 48 states of the U.S., southern Canada, northern Mexico, and the contiguous ocean areas on a 1:20 million scale. This numerical analysis duplicates as closely as possible the manual analysis. A further description of this chart is contained in the National Weather Service Forecasters Handbook WBFH No. 1.

FNWC Monterey provides a 36-hour pressure height and temperature prognosis for the 500-, 400-, 200-, and 100-mb levels. To obtain this chart the 500- to 400-mb temperature lapse rate is extrapolated upward to its intersection with the 200- to 100-mb lapse rate extrapolated downward. The height at which this intersection takes place is taken as the tropopause height. (See fig. 7-30.)

By this method the tropopause is obtained as a continuous surface. Its actual leaf structure is not portrayed. The tropopause height can be used as an approximation to the level of maximum winds.

SUPPLEMENTARY UPPER AIR ANALYSIS

The basic upper air analysis is the constant pressure analysis. In conjunction with this basic analysis it is sometimes necessary, and at all times beneficial, to conduct concurrent supplementary types of analyses of upper air properties in order that the fullest use be made of

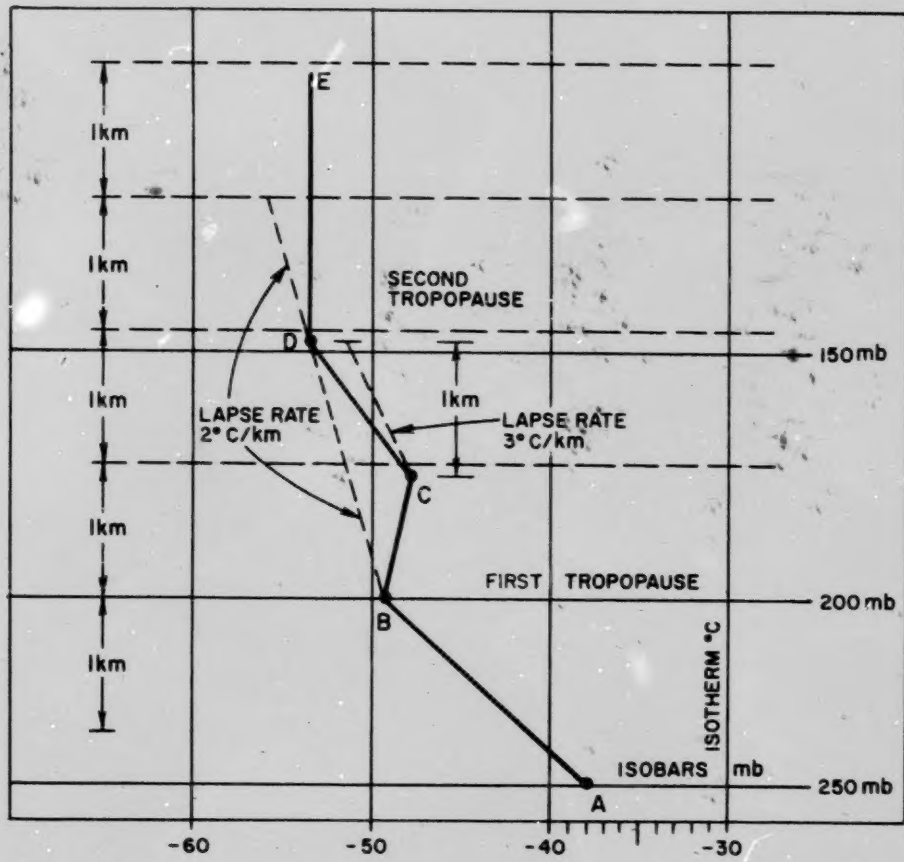

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Figure 7-29.—Tropopause determinations based on WMO definitions.

upper air information to lead to the end product, the forecast. Most of these charts are constructed from either reported or derived data from upper wind and upper air reports. It is not feasible to list or explain all the types of upper air charts currently being produced by the National Meteorological Center. Only the space differential (thickness) analysis, time differential, and advection charts are covered here.

SPACE DIFFERENTIAL (THICKNESS) ANALYSIS

Construction

Space differential analysis is an analysis involving the utilization of the contours from two analyzed constant pressure surfaces whereby the contours (height values) of the lower level are subtracted graphically from the contours (height values) of the upper level (as illustrated in fig. 7-31) to obtain the "differential" (thickness) between the levels.

The thickness between the levels (thickness charts may be constructed between any two pressure surfaces) indicates the mean temperature of that layer; the amount and type of advection occurring (thermal wind); the density of the layer; deepening and filling of pressure systems; the vertical wind shear; and cyclogenesis. Space differential charts are used primarily as a basis for advection charts; in forecasting sea level pressure; in forecasting upper heights; in forecasting deepening and filling of

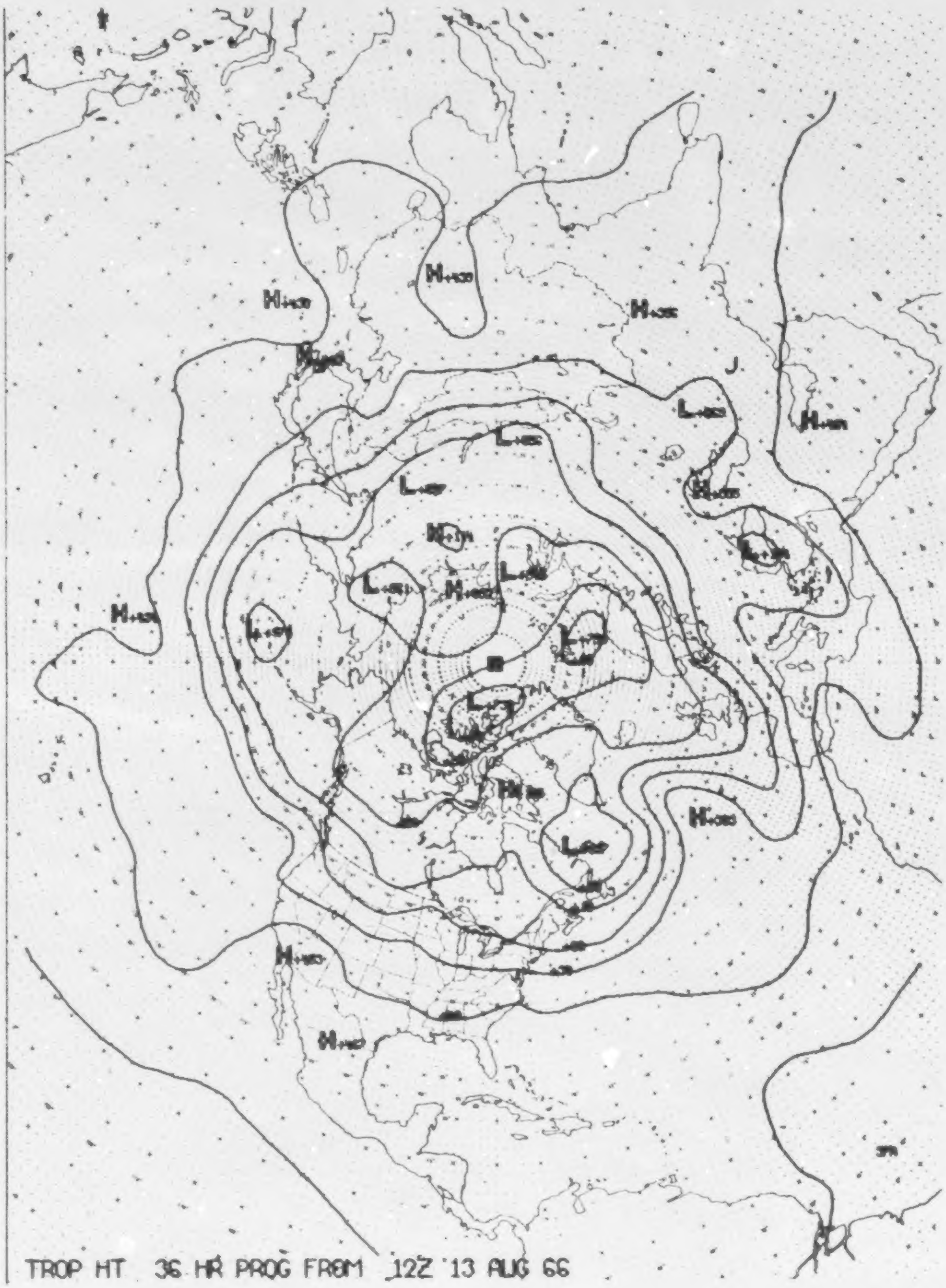
TROP HT 36 HR PROG FROM 12Z 13 AUG 66

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Figure 7-30.—Tropopause height 36-hour prognosis.

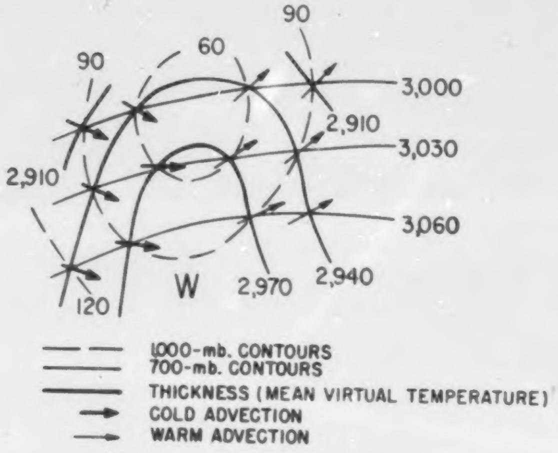

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Figure 7-31.—Graphical subtraction of contours (heights in meters).

pressure systems; in forecasting (in conjunction with other things) the direction and speed of motion of surface and upper air pressure systems; and in checking the consistency of frontal analysis. The most commonly used differential charts are: the 1,000- to 700-mb thickness chart, the 1,000- to 500-mb thickness chart, the 700- to 500-mb thickness chart, and the 500- to 200-mb thickness chart.

In constructing thickness charts it is best to use acetates; next best is to use a clean chart. First, trace the lower-level contours onto either the chart or acetate, and label appropriately. The lower-level contours should have a distinctive color (usually yellow). Next, trace the upper-level contours onto the chart or acetate, also in a distinctive color (usually green). This method eliminates lines, such as isotherms in the analysis area. Then, graphically subtract the lower contours from the upper and connect the points of equal thickness with a dashed black line, and label them.

To start the analysis, it may be helpful to determine arithmetically the height difference at a few points. As experience is gained by the Aerographer's Mate, enough proficiency is acquired so that the hand holding the pencil is able to follow the eye as it sees the points. Thickness lines must cross the contours of both charts in the same linear direction. They cross from greater heights to lower heights or from lower heights to greater heights on BOTH charts at any given intersection. Thickness lines cross contours of either chart only at intersections and do not cross each other.

Two consecutive contours of either constant pressure chart must be separated by a contour of the other constant pressure chart or by a thickness line. Two consecutive thickness lines must be separated by a contour of one of the constant pressure charts. A thickness line between two intersections is proportional to the thermal wind. The thermal wind is the mean wind shear vector in geostrophic balance with the gradient of mean temperature bounded by two isobaric surfaces. The thermal wind is directed along isotherms with cold air to the left in the Northern Hemisphere. It can be shown vectorially by subtracting the wind vector of the lower isobaric surface from the wind vector at the upper isobaric surface as illustrated in figure 7-3.

Uses of the 1,000-500 mb Thickness Chart

The 1,000-500 mb thickness chart is one of the primary tools used at the National Meteorological Center and can be equally as important to you. Some of the important relationships between thickness and between fronts and cyclones in the various stages of development are given in the following sections. Figure 7-32 illustrates most of the important details of thickness patterns in relation to fronts.

The most important features stressed and used are:

1. Warm sectors are relatively homogenous; the concentration of thickness contours is on the cold side of frontal systems. This will be better defined with moderate to strong fronts.
2. The cold trough in the thickness contours lies to the rear of the surface depression, halfway between the surface depression and the next upstream surface ridge, or high.
3. Thickness contours are anticyclonically curved in advance of warm fronts and cyclonically curved behind cold fronts.

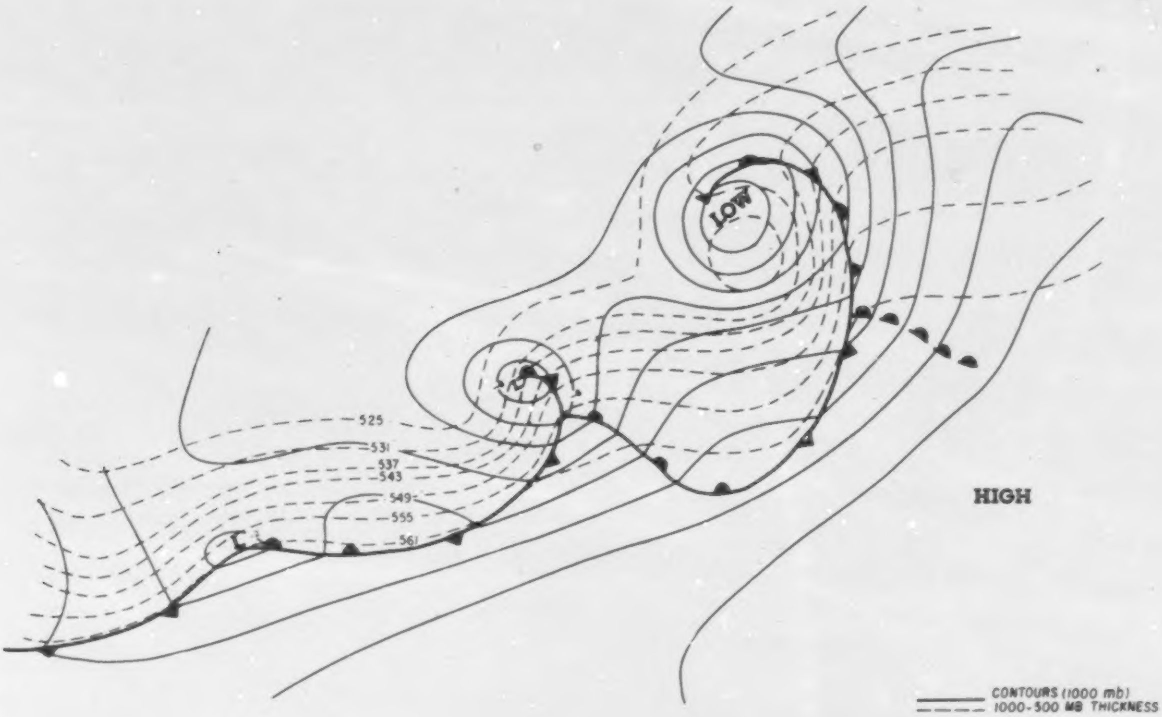

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Figure 7-32.–Illustration of the relation of thickness patterns to fronts.

4. The spacing of thickness contours behind cold fronts is closer than in advance of warm fronts, reflecting the generally accepted idea that cold fronts are more intense than warm fronts.

5. The distance between the thickness jet which is usually located horizontally in the same position as the 500-mb jet and cold fronts is less than with warm fronts, reflecting the steeper slope of cold fronts as compared with warm fronts.

Adherence to these important features of the thickness model insures the proper slope of systems between 1,000 and 500 mb; the proper relationship between surface fronts and the polar jetstream; and surface frontal analyses that portrays a meaningful picture of the 3-dimensional temperature structure.

ADVECTION CHARTS

Advection charts are essentially a part of space differential charts. They are usually constructed on the space differential chart, although a separate chart may be used to indicate the advection patterns. Advection patterns are indicated at every point of intersection of the constant pressure charts. The advective flow is nearly perpendicular to the thickness lines and is indicated with a single-shaft, single-barb blue arrow for cold advection and a single-shaft, single-barb red arrow for warm advection. Advection arrows indicate the mean direction of flow between the two isobaric surfaces. If the arrow crosses the thickness lines from higher value to lower value, the advection is warm. If the arrow crosses the thickness lines from lower values to higher values, the advection is cold. Advection charts may be constructed between any two desired isobaric surfaces.

Rules of thumb for the Aerographer's Mate to follow when constructing advection arrows are: when the lower contour is to the left while looking downstream, advection is cold; when the lower contour is to the right while looking downstream, advection is warm. Refer to figure 7-31 for an illustration of the proper method of constructing advection arrows.

A derivative of the thermal wind equation and the advection arrows is the thermal wind rule. This rule simply states that in the Northern Hemisphere if the wind backs with height, cold advection is indicated, and if the wind veers with height, warm advection is indicated. That this rule is true is shown in figure 7-33.

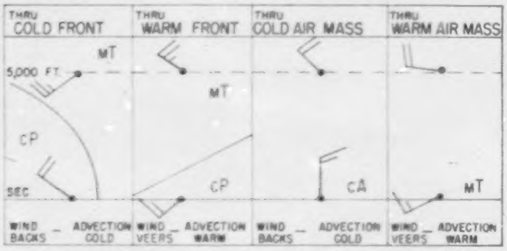
THRU COLD FRONT
THRU WARM FRONT
THRU COLD AIR MASS
THRU WARM AIR MASS
mT
5,000 FT
mT
cP
SFC
cP
cA
mT
WIND BACKS _ ADVECTION COLD
WIND VEERS _ ADVECTION WARM
WIND BACKS _ ADVECTION COLD
WIND VEERS _ ADVECTION WARM

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Figure 7-33.—The thermal wind rule (Northern Hemisphere).

TIME DIFFERENTIAL CHARTS

Time differential charts serve the main purpose of tracking rise and fall centers on upper air charts (usually the 500-mb chart), although other levels may be chosen for this purpose as well. These charts are usually drawn at 24-hr intervals in order to minimize diurnal effects which might otherwise invalidate some of the features of these charts.

The time differential charts are constructed in the following manner: take two charts, 24 hours apart, of the same isobaric surface (500 mb). Place the older chart on the bottom, and a clean acetate or blank chart over both. Over a light table construct lines of equal difference at each of the intersections of the two charts much in the manner used to construct the thickness charts. Start the analysis by algebraically subtracting the new contour values from the old contour values and connecting the points of equal difference. Areas of falling heights are indicated by drawing the height change lines as solid red lines; areas of rising heights are indicated by drawing the height change lines as solid blue lines, drawing the zero height change line as a purple line. The centers of rising and falling heights are then transposed onto a blank chart used solely for this purpose. Succeeding positions of the rise and fall centers are connected in the same manner as are pressure system centers on the constant pressure charts.

The individual height change lines should be marked in the manner of the contours on constant pressure charts. The center values of the rise and fall centers should be entered on the tracking chart. For an illustration of the method of constructing time differential charts, see figure 7-34.

COMPUTER PRODUCTS

Space differential (thickness), and advection charts are available in the form of computer derived products.

The 1,000 to 700, 1,000 to 500, 700 to 500, and 500 to 200 mb thickness (differential) charts are all available in computer form. As discussed earlier, computer derived vorticity advection chart are available for utilization in the analysis process. Time differential charts may be constructed by utilizing a series of thickness charts and are therefore not available in chart form over computer circuits.

SUMMARY OF UPPER AIR ANALYSIS RULES

Observation of many analyses by inexperienced personnel reveals some typical and often-repeated errors in upper-air analysis. The following list is by no means all inclusive.

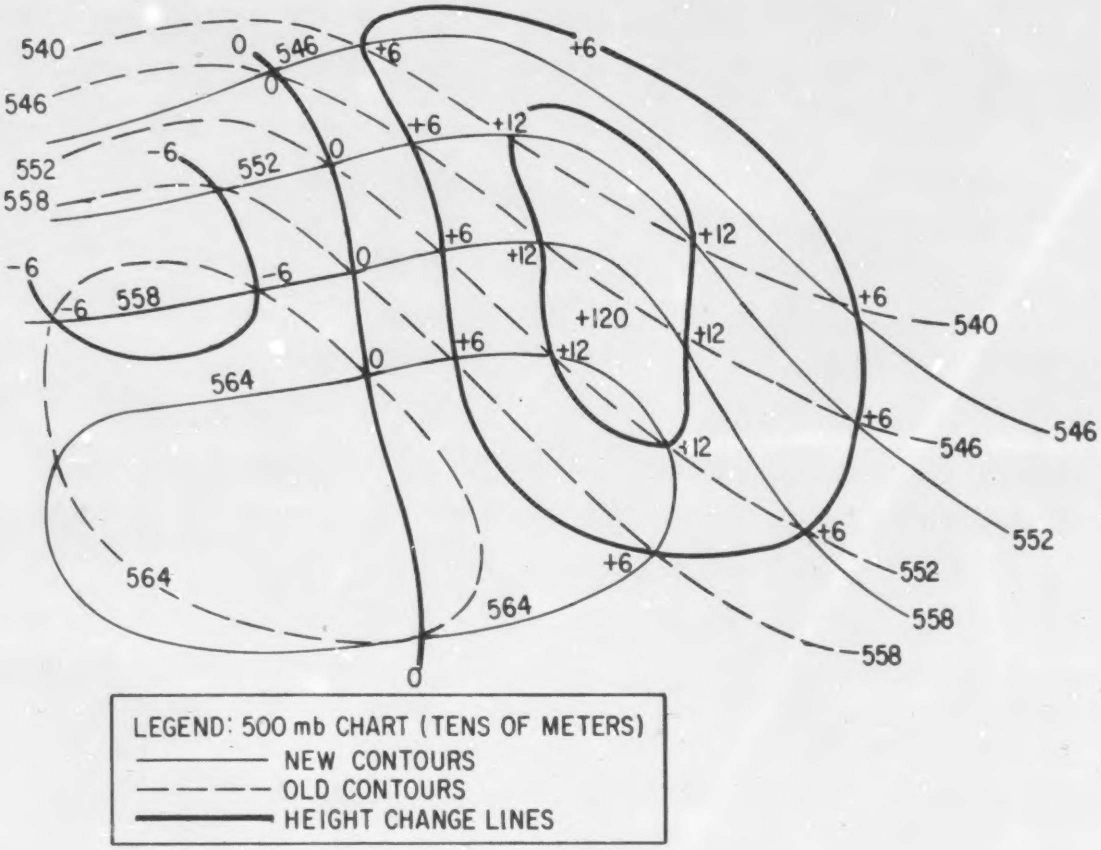


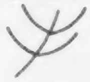
Figure 7-34.–Time differential chart.

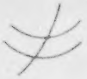
1. Contours (isohypses) are to be drawn parallel to winds, where possible. This will not always be possible due to errors in observations and nongeostrophic and/or nongradient winds. Always check the plotted windshaft with the number indicating direction before deciding the observed wind is impossible to draw to.

2. Use history. Troughs, ridges, highs and lows may change slightly their configuration and orientation and have differential movement. History tells what these features looked like 12 and 24 hours ago and in what general area to start looking for them on the current map. Always check the previous analyzed map before starting analysis.

3. Wind direction cannot change discontinuously along a contour.

4. In the vicinity of centers (highs and lows) do not stop analysis merely because there are no actual data. Use wind scales, history chart and common sense to get central contour height and its position. If an intermediate contour helps to define location of a center, put it in (with a dashed line).

5. Do not overemphasize the cyclonic curvature at troughs in the form of a kink in the contour (i.e. ). This is absolutely wrong. It implies that the trough is a front.

Instead do this:  Note the relative maximum of cyclonic curvature at the trough.

6. Do not throw away data at first glance. Over the ocean area the following reports are excellent, with the list in descending order of "level of belief".

a. Permanent ships, as weather ships November, Papa and others; traveling commercial and military surface craft; also, permanent land stations as the Hawaiian group, Midway, etc.; all of the foregoing are "on-time" reports;

b. Weather reconnaissance (recon) reports; these are plotted about a □ with T, T_d, 500-mb height and usually wind; and

c. Commercial and military aircraft reporting flight-level temperature, wind, and sometimes a "D" value; extrapolations to 500-mb level must then be made.

7. With weather recons and other aircraft reports, time of observation must be taken into account. Correct the data to map time by time interpolation or extrapolation.

8. Over the ocean, continuity and history are as important as any observation or extrapolation. Do not lose systems for lack of data over the ocean.

9. Troughs and centers moving 50 kt or more are to be doubted unless overwhelming evidence indicates this to be true. Movement of troughs and centers embedded in the westerlies is generally W-E with some meridional component. The cold lows and warm highs tend to move slowest, (usually less than 0.5° lat/hr). On occasion the centers of these dynamic cold and warm systems move westward with some meridional component. At times they are stationary.

10. Considerable trouble is experienced in northern Canada due to lack of data. This is an area where one must analyze, not just draw lines. Most troubles could be cleared up by observing accuracy of contour analysis (use Buys-Ballot's law), by drawing in a few intermediate contours, and by using the geostrophic wind scale.

11. In using the geostrophic wind scale to determine correct spacing of isohypses on upper-air charts, the spacing must be less than the observed wind indicates where the curvature of the isohypses is cyclonic; the converse is true for anticyclonic circulation. This is only another way of saying that winds in cyclonic regions are subgeostrophic, in anticyclonic regions supergeostrophic.

12. In isotherm analysis keep in mind there is a tendency for the lows (highs) at 500 mb to be cold (warm) with a definite closed isotherm nearly coincident with the pressure center. Long-wave troughs are cold, ridges warm, with reverse relation for short waves.

13. Jagged, ragged or nervous lines lend an unprofessional appearance to an analyzed map. Avoid discontinuities in the isolines.

14. Short-wave troughs move through long-wave patterns; hence their orientation and movement traces out long wave patterns. Troughs are normally concave to the direction from which they are moving. Check history map for most recent configuration. Marked changes in configuration should be doubted.

15. Check vertical consistency by reference to analyses at adjoining levels, especially the lower levels:

a. Troughs and lows must slope toward coldest air.

b. Ridges and highs must slope toward warmest air.

c. Cold lows have little or no slope.

d. Warm lows at low levels become short-wave (warm) troughs aloft.

e. Wave cyclones at sea level become short waves aloft.

f. Occluded cyclones become cold or cut-off lows aloft.

CHAPTER 8

FORECASTING UPPER AIR SYSTEMS

To prepare prognostic charts, both surface and upper air, we must first make predictions of the various weather systems found on each type chart. Inasmuch as the current surface and upper air charts reveal in detail the current state of the weather, so should the prognostic charts reveal accurately and in detail the future state of the weather. Constructing prognostic charts is no easy task; nevertheless it is not impossible.

The ideal approach to preparing upper air and surface prognostic charts dictates that the Aerographer's Mate first begin with the upper levels and then translate the prog downward in terms of a surface prog. The two are so interrelated that consideration of the elements on one should not be made independently of the other.

Prognostic charts are constructed at the National Meteorological Center (by Numerical Weather Prediction methods) and Fleet Numerical Weather Facility, Monterey. The resultant products are transmitted over their respective facsimile networks.

Overseas Fleet Weather Centrals and Facilities also construct and transmit progs. We are all too often inclined to rely solely on these data with few or no actual computations made by ourselves. Since these prognostics are generally for large areas, this procedure could lead to many erroneous forecasts and mental lethargy on the part of the Aerographer's Mate.

For these and other reasons it is important that the Aerographer's Mate not only understand the methods by which prognostic charts are constructed but their limitations as well. In this chapter we will discuss some of the more common methods and rules for forecasting upper air features. In the following chapter, methods and techniques for progging the surface chart will be considered. These methods can be used in constructing your own prognostic charts where data are not available from other sources and/or to check on the prognostic charts made by other sources.

GENERAL PROGNOSTIC CONSIDERATIONS

It is incumbent upon the forecaster to consider all applicable rules, draw upon experience, and consult any valid available objective aids to produce the best possible forecast from data that are available. This evolves into both a subjective and objective approach to the prognosis and the forecast.

A forecaster has examined many aspects of the weather picture from both surface and upper air maps by the time he issues his forecast. Some conditions are overlooked or soft-pedaled, while others are emphasized. The forecaster must depend heavily upon his experience and knowledge with past history of the weather because similar analogous conditions yield similar consequences. Some forecasters may decide to overlook a clearly defined parameter, such as surface pressure, because through his experience, or the experience of others, he may decide that it is not a decisive factor. In this case he may concentrate on the general trend of upper wind, the humidity and stability revealed by soundings, and other factors. The mental processes involved in making the prognosis and the forecast are also guided by the forecast requirements.

An objective system of forecasting certain atmospheric parameters can parallel and often exceed the skill of an experienced forecaster. An objective system also often eliminates the uncomfortable moments of hesitation that every forecaster has undoubtedly suffered when he has been pressed for a forecast during nondescript conditions. However, the objective process should not necessarily take precedence over the subjective method, but rather the two should be used together, where and when applicable, to arrive at the most accurate forecast possible in the shortest period of time. At any rate, whichever system is used, some systematic method should be adopted, whereby all of the factors possible are cataloged for ready reference as a checkoff list.

HAND DRAWN ANALYSIS

The correct methods and procedures to be utilized in analyzing upper air charts have been covered in Chapter 7. A properly drawn hand analysis provides the forecaster with the basic and most important tool in constructing an upper air prognostic chart which will attain a high degree of validity. Such information as windspeed direction, temperature, dew point depression, and heights is readily available for the forecaster to integrate into the various objective methods of producing a prognostic chart.

COMPUTER PRODUCTS

The Fleet Numerical Weather Central presently provides a great variety of charts for dissemination to shore and fleet weather units. These include analysis and prognostic charts ranging from subsurface oceanographic charts to the depiction of the troposphere, as well as a number of specialized charts. A number of these charts have been discussed in the preceding chapter. A complete listing of the various charts is contained in the Naval Weather Service Computer Products Manual, NavAir 50-1G-522.

APPLICATION OF SATELLITE CLOUD PHOTOGRAPHS

As a further aid, satellite pictures, mosaics, and nephanalysis can also be used in preparing a prognostic chart. The availability of satellite data will be extremely varied. Some stations will receive only teletype or facsimile data while others will be equipped to receive direct readout of satellite photographs.

OBJECTIVE FORECASTING TECHNIQUES

The use of objective forecasting techniques in preparation of upper air prognostic charts affords the forecaster a better opportunity to arrive at a more accurate chart. Experience in itself is not enough to forecast the movement or intensity of upper air systems, but coupled with basic objective techniques provides a sound basis for the forecaster to prepare an accurate and valid chart and forecast.

FORECASTING THE MOVEMENT OF TROUGHS AND RIDGES

The techniques covered in this section apply primarily to long waves. Many of them are applicable to short waves as well and it will be stated when and where they are applicable. A long wave is by definition a wave in the major belt of westerlies which is characterized by large length and significant amplitude. Therefore, the first step in progging long waves is to determine their limits. Large closed lows and highs, cutoff systems, and the like must be temporarily set aside for separate consideration. This done, we are ready to prog the movement of the long waves. There are several basic approaches to the progging of both long and short waves. Chiefly these are extrapolation, CAVT's, Petterssen's Wave Speed Nomogram, the Grid Method, isotherm-contour relationship, and the location of the jet maximum in relation to the current in which it lies.

Extrapolation

Knowledge of the past history of the systems affecting the area of interest is fundamental to the success of forecasting. Experience has shown that most atmospheric systems usually change slowly and continuously with time. That is, a continuity in the weather pattern is exhibited in a sequence of weather charts. When a particular pressure system or distribution exhibits a tendency to continue without much change it is said to be persistent. These concepts of persistence and continuity are very useful forecast aids.

The extrapolation procedures used in forecasting may vary from simple linear extrapolation to the use of more complex mathematical equations and analog methods based on theory or long experience. Two such methods are presented here: one is simple extrapolation, the other a more complex method based on kinematic extrapolation.

The practicing forecaster usually extrapolates past and present conditions to obtain future conditions in accordance with scientific principles and years of practical experience.

SIMPLE EXTRAPOLATION.–The simplest method of forecasting both long and short wave movement is that of extrapolation.

Simple extrapolation is merely the displacement of the trough or ridge to a future position based on past and current movement and expected trends. It is based on the assumption that the changes in velocity of the pressure patterns are slow and gradual. However, there are pitfalls to this method in that new developments frequently occur which were not revealed from present or past indications upon which the extrapolation was based. However, if such developments can be forecast by other techniques, allowance can be made for them.

Extrapolation for short periods on short waves is generally valid. The major disadvantage of extrapolating the movement of long waves, or long period movements of short waves, is that past and present trends do not continue indefinitely. This can be seen when we consider a wave with a history of retrogression. The retrogression will not continue indefinitely, and we must look for indications of its reversal, that is, progressive movement.

Constant Absolute Vorticity Trajectories (CAVT's)

This method is used to determine the movement of long wave troughs and ridges at the 500-mb level.

To compute CAVT's the following data at the inflection point is required: maximum wind speed, wind direction (angle with east), and latitude value to be used in the CAVT table.

The inflection point is that point along each ridge or trough where the relative vorticity changes from positive to negative, or vice versa. In other words, the actual curvature of the contour changes from cyclonic to anticyclonic, or vice versa. In this area, the point with zero wind shear (strongest windspeed) is designated as the CAVT inflection point. The point will always be upstream from the trough or ridge concerned. (See fig. 8-1.)

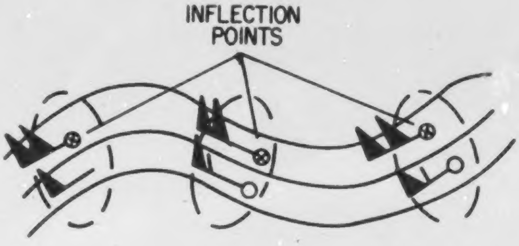


Figure 8-1.–Selection of the inflection point for CAVT's.

Information regarding procedure to be followed in using the CAVT method of forecasting the movement of troughs and ridges may be found in various publications which are listed in Navy Weather Research Facility Reports and Publications (NWRF 00-0369-143) and updated revisions.

Petterssen's Nomogram

Petterssen's wave speed nomogram utilizes the wavelength (L) in degrees latitude or nautical miles; the windspeed at the core of the current at the trough or ridge line (U); the distance from the core of the current to location of U where the windspeed is one-half of U in the same units as L, (B) which provides the value for L/B; the wave axis of inclination from the meridian of the trough Υ (gamma); and the latitude Φ (phi) where U is determined.

Using the data given in figure 8-2, enter the nomogram (fig. 8-3) at the top under the value of 3,000 nautical miles and go downward to the

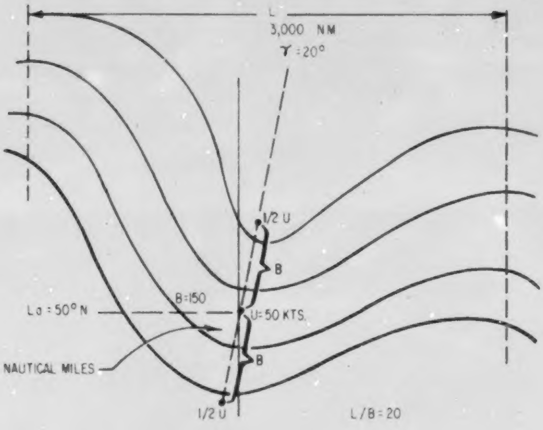


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Figure 8-2.—A long wave situation for use with Petterssen's wave speed nomogram.

value of Υ (gamma). From the point of intersection of Υ (gamma), go horizontally across to the value of Φ (phi), and thence vertically downward to the value of U. From the point where you intersected U, go horizontally across the nomogram to the value of L/B and thence vertically upward until you intersect the slanting line which gives the movement of the long wave in knots or degrees latitude per day. Your result in this case is 2° lat/day. Other waves should be similarly evaluated.

The major obstacles to successful use of the nomogram are that L and Υ (gamma) are not always definitive, and that U and B, and therefore L/B are not always accurate. Despite these drawbacks, the nomogram is of value in that it does yield a rough indication of both the direction and extent of movement and this is of value, because the Aerographer's Mate can then smooth out the result.

The best results from the use of this technique will be obtained if the troughs and ridges to be computed are selected under the following conditions: Adequate upper wind data are necessary; only cases should be selected where the streamlines have a well defined sinusoidal pattern from the trough or ridge in question for a distance L/2 upstream. Symmetry of wavelengths upstream and downstream about the trough or ridge is not required; a simple and fairly symmetric velocity profile along the trough or ridge line is necessary.

A test of 158 troughs and ridges indicated the following corrections should be applied to the results of the computations:

1. Deduct 1 longitude degree per day from each computation of trough speed.
2. Deduct 3 longitude degrees per day from the computed motion of each ridge east of the Rocky Mountains.
3. Deduct 5 degrees per day from the computed motion of each ridge over the Rockies.

Grid Method

This method is used for forecasting the movement of short wave troughs and closed lows at the 500-mb level. A grid is constructed as a 20-degree latitude square, true at latitude 45 degrees for the type map projection used. The displacment of the low or trough is given by the mean geostrophic wind over the area covered by the grid. (See fig. 8-4 for example of grid.)

First we will discuss the movement of 500-mb short wave troughs for a 24-hour forecast period.

1. Aline the grid so that the centerline DOC is over the trough with point "O" approximately in the middle of the trough.
2. Read off height points from the 500-mb height field at all points from A to F (in this example meters are used).
3. Using the following formula, substitute the heights for the letters in the equation. Note plus values indicate east movement, minus values west movement.

"X" [(A-B) + (C-D) + (E-F)] = E-W MVT,

where E = east.

The solution to the equation should be positive to indicate eastward movement, since short waves are progressive. Movement is in degrees of latitude. "X" is the latitude factor using latitude at point "O". See Table 8-1 for the "X" factors for two different map projections and the values for both feet and meters. (Feet using hundreds of feet, and meters using tens of meters.)

For moving short wave troughs, a systematic correction of minus 1 degree of latitude per

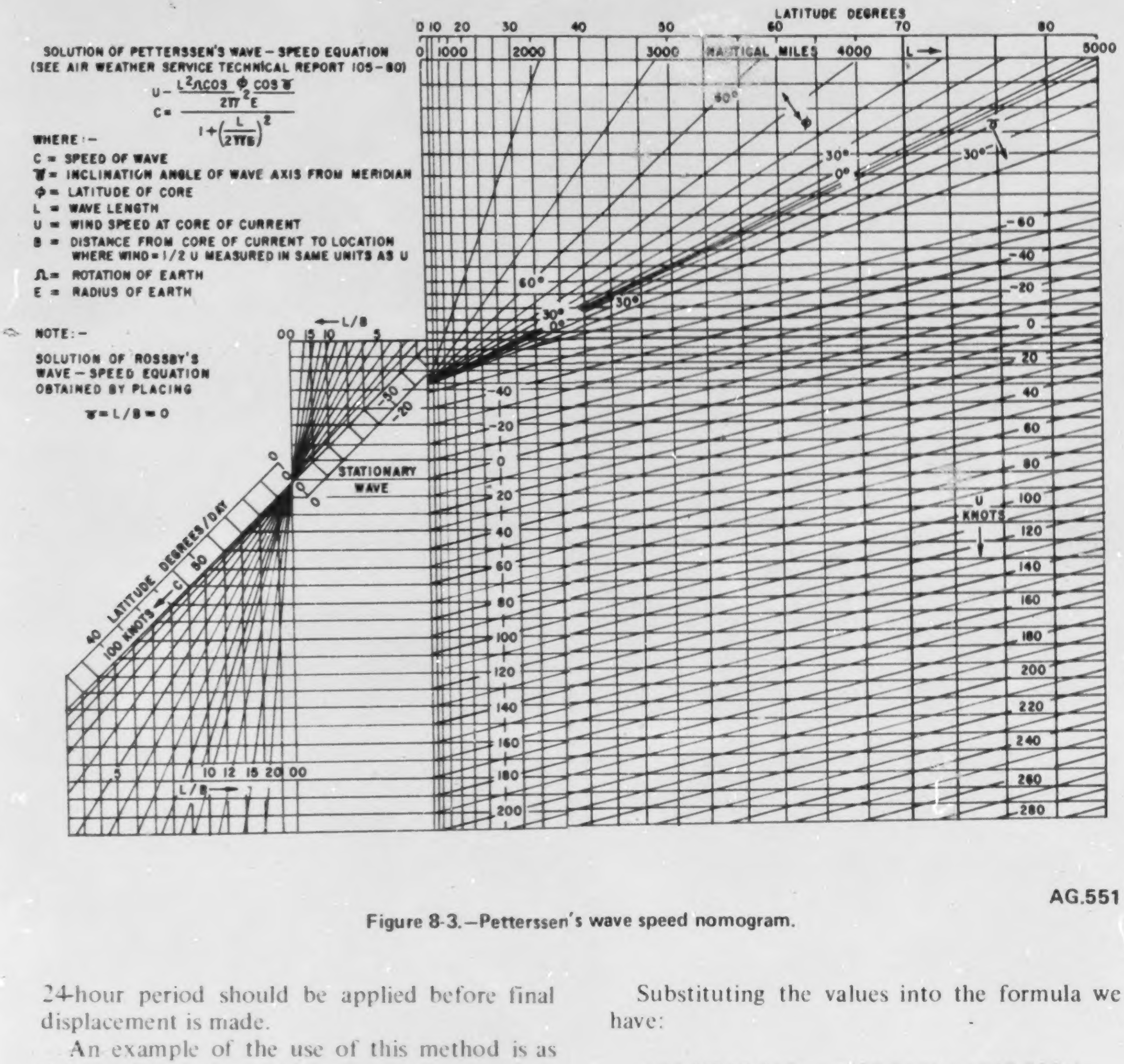

Figure 8-3.—Petterssen's wave speed nomogram.

24-hour period should be applied before final displacement is made.

An example of the use of this method is as follows. Suppose we had constructed our grid on a polar stereographic chart and read off the following values for the points indicated:

Latitude at point O = 40 degrees
Height at point A = 5,430 meters
Height at point B = 5,120 meters
Height at point C = 5,550 meters
Height at point D = 5,150 meters
Height at point E = 5,670 meters
Height at point F = 5,330 meters

Substituting the values into the formula we have:

$$.125\ [(543-512) + (555-515) + (567-533)]$$
$$.125\ [(31) + (40) + (34)]$$
$$.125 \times 105 = 13.13 - 1.00 = 12.13^\circ \text{ lat per day}$$

Normally good results are obtained when working with troughs of small amplitude where amplitude decreases or remains constant with time. However, overforecasting is evident when this method is applied to troughs which deepen rapidly. As a rough guide, this method should

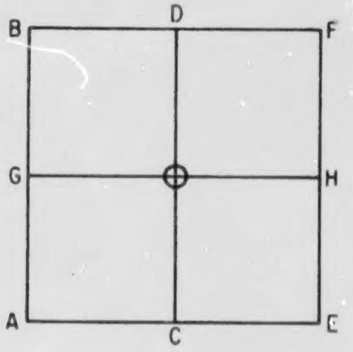

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Figure 8-4.—500-mb grid.

Table 8-1.—"X" factors for use with the grid equation.

	Lambert Conformal		Polar Stereographic	
Latitude	Feet	Meters	Feet	Meters
25	.57	.187	.67	.220
30	.48	.157	.54	.177
35	.42	.138	.45	.148
40	.37	.121	.38	.125
45	.34	.112	.34	.112
50	.31	.102	.30	.098
55	.30	.098	.27	.089
60	.28	.092	.25	.082
70	.26	.085	.23	.075

not be used when the contours in the neighborhood of point "O" (initial point) greatly exceed an amplitude of 15 to 20 degrees of latitude.

Isotherm-Contour Relationships

The forecaster should always examine the long waves for the isotherm-contour relationships and then apply the rules for the movement of long waves. These rules were stated in chapter 4 of this training manual. The rules do not yield a very quantitative movement, but if the rules confirm the CAVT results or the nomogram results, or both, they have served the purpose of confirming previous results. This can be seen in a situation where CAVT's yield small progressive movement and the nomogram yields a small retrograde movement. If the isotherm-contour relationship confirms retrograde motion, we are justified in placing more faith in the nomogram result. A number of observations and rules are stated regarding the progression, stationary characteristics, or retrogression of long waves. These rules are stated in the following section.

PROGRESSION OF LONG WAVES.—Progression (eastward movement) of the long waves is usually found in association with relatively short wave lengths and well defined major troughs and ridges in the middle and upper troposphere. At the surface, there are usually only one or two prominent cyclones associated with each major trough aloft. Under the forward part of each major ridge there is usually a well developed surface anticyclone moving toward the east or southeast. The 24-hour height changes at upper levels usually have a one-to-one association with major troughs and ridges (motion of maximum height fall and rise areas associated respectively with major trough and ridge motion). The tracks of the height change centers depend on the movement and changes of intensity of the long waves, and often seem somewhat erratic.

STATIONARY LONG WAVE PATTERNS.—Once established, stationary long wave patterns usually persist for a number of days. The upper air flow associated with the long wave pattern constitutes a steering pattern for the smaller scale disturbances. These small scale troughs and ridges, with their associated heightchange patterns and weak surface systems, move along in the flow of the large scale, long wave pattern. Minor troughs intensify as they move through the troughs of the long waves and weaken as they move through the ridges of the long waves. The same changes in intensity occur in those sea level troughs or pressure centers which are associated with minor troughs. Partly as a result of the presence of these smaller scale systems, the troughs and ridges of the stationary long waves are often spread out and hard to locate exactly.

RETROGRESSION OF LONG WAVES.–A continuous retrogression of long wave troughs, in which the wave troughs are conserved, is a rare event. The usual type of retrogression takes place in a discontinuous fashion whereby a major trough weakens, accelerates eastward, and is transformed into a minor trough while a major wave trough forms to the west of the former position of the old one. New major troughs are generally formed by the transformation of minor troughs into deep cold troughs. Since old troughs are weakening and new ones are developing during this process, it would be improper to apply any wave velocity equation to this type situation.

Retrogression is seldom a localized phenomenon, but as a rule appears to occur as a series of retrogressions in several long waves. Retrogression generally begins in a quasi-stationary long wave train when the stationary wavelength shows a significant decrease. This can happen as a result of a decrease in zonal wind speed, or of a southward shift in the zonal westerlies. Some characteristics of retrogression are as follows. Trajectories of 24-hour height change patterns at 500-mb deviate from the band of maximum wind. New centers appear or existing ones rapidly increase in intensity. Rapid intensification of surface cyclones occurs to the west of existing major trough positions.

LOCATION OF THE JET STREAM.–In chapter 4 of this training manual we discussed the migration of the jetstream both northward and southward. Some general considerations can be made concerning this migration and the movement of waves in the troposphere.

In a northward migrating jet, a west wind maximum emerges from the tropics and gradually moves through the lower midlatitudes. Another maximum, initially located in the upper midlatitudes, advances toward the Arctic Circle while weakening. Open progressive wave patterns with pronounced amplitude and a decrease in the number of waves due to cutoff centers exist. The jet is well organized and troughs extend into low latitudes.

As the jet progresses northward, the amplitude of the long waves decrease and the cutoff lows south of the westerlies dissipate. By the time the jet reaches midlatitudes a classical high index situation exists. Too, we have weak, long waves of large wavelength and small amplitude, slowly progressive or stationary. Few extensions of troughs into the low latitudes are present and in this situation the jetstream is weak and disorganized.

As the jet proceeds farther northward, there will often be a sharp break of high index with rapidly increasing amplitude of the flow aloft. Long waves retrograde. As the jet reaches the upper midlatitudes and into the sub-Arctic region, it is still the dominant feature, while a new jet of the westerlies gradually begins to form in the subtropical regions. Long waves now begin to increase in number and there is a reappearance of troughs in the Tropics. The cycle then begins over again and repeats itself.

With a southward migrating jet, the processes are reversed from that of the northward moving jet. It should also be remembered that short waves are associated with the jet maximum and move with about the same speed as these jet maximums.

FORECASTING THE INTENSITY OF TROUGHS AND RIDGES

Forecasting the intensity of long waves often yields nothing more than a sign of the expected intensity–greater than or less than present intensity. For instance, if deepening or filling is indicated, but the extent of deepening or filling is not definite, the Aerographer's Mate is forced to rely on experience and intuition in order to prog the amount of deepening or filling. Numerical Weather Prediction Methods do forecast the intensity of upper waves with a great deal of success. If available, you should check your intensity and movement predictions against these prognoses.

Extrapolation

Patterns of upper level charts are more persistent than those on the surface. Therefore, extrapolation gives better results on the upper air charts than on surface charts. In using pressure changes aloft, the procedure is to extrapolate height change centers and add the changes to the current height values in order to get the prognostic map. Changes in the shape of the upper air pattern may be forecast by this method, whereas pure extrapolation would not indicate these variations.

Use of Time Differentials

The time differential chart is discussed in chapter 7 of this training manual.

The time differential chart for the 500-mb level shows the history of what changes have taken place at the 500-mb level at 24-hour intervals. In considering the information on the time differential chart, those centers with a well defined history of movement will be of greatest importance. Take into consideration not only the amount of movement, but also the changes in intensity of the centers. Single centers with no apparent history should be treated with caution especially with regard to their direction of movement which is usually in a downstream position from their current position. The information as indicated on the time differential chart should be used to supplement the information already obtained from previous considerations and when in agreement can be used as a guide for the amount of changes to the prognostic contours in a given area.

Normally, the 24-hour height rise areas can be moved with the speed of the associated short wave ridges and the speed of the fall centers with the [illegible]d of the associated wave troughs. Either k[illegible] the intensity of the height change centers unchanged or modify them according to past developments. It must be remembered that height change centers may be present due to convergence or divergence factors and may not have a short wave trough or ridge associated with them. However, normally with short wave indications, a change center will appear and move in the direction of the contour flow. Be cautious not to move a height change center with the contour flow if it is due primarily to convergence or divergence.

Once you have progged the movement of the height change centers and determined their magnitude, apply the change indicated to the height on the current 500-mb chart and use these points as guides in constructing prognostic contours.

Consideration of CAVT's

As mentioned earlier, by comparing the amplitude of the troughs and ridges on the current chart and the amplitude as determined by the CAVT tables, intensification or weakening can be determined.

Isotherm-Contour Relationship

In long waves, deepening troughs are associated with cold advection to the west side of the trough and filling troughs with warm advection. The converse is true for ridges. Warm air advection into the western side of a ridge indicates intensification and cold air advection indicates weakening. This rule is least applicable east of the Continental Divide and probably east of any high mountain range where winds are from the west. In short waves, deepening short wave troughs have cold air advection into the west side of the trough, particularly if a jet maximum is in the northerly current of the trough and filling is indicated by warm air advection on the western side.

In the above paragraph, the advection is not the cause of the intensity changes, but rather is a "sign" of what is occurring. High level convergence/divergence is the cause.

Effect of Super and Subgradient Winds

Figure 8-5 (A) through (D) shows the effect of the location of maximum winds on the intensity of troughs and ridges.

Explanation of figure 8-5 is as follows:

1. When the strongest winds aloft are the westerlies on the western side of the trough, the trough deepens (fig. 8-5(A)).
2. When the strongest winds aloft are the westerlies in the southern quadrant of the trough, the trough moves rapidly eastward and does not change in intensity (fig. 8-5(B)).
3. When the strongest winds are the southwesterlies between the trough and the downstream ridge, the trough decreases in intensity (fig. 8-5(C)).
4. Sharply curved ridges with excessive contour gradients are unstable and rotate rapidly clockwise, causing large height rises and filling in

the trough area downstream and large height falls in the left side of the strong gradient ridge (fig. 8-5(D)).

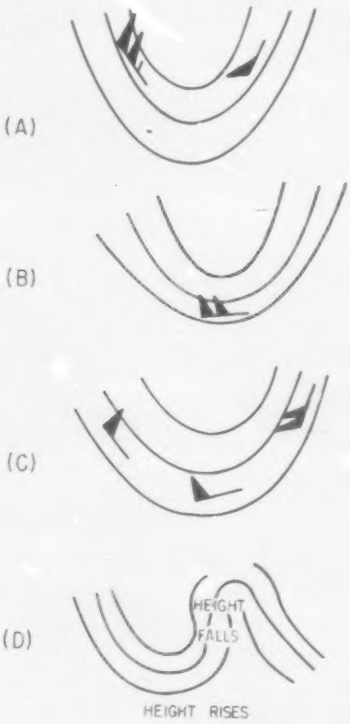

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Figure 8-5.–Effect of super and subgradient winds on the deepening and filling of troughs. (A) Strongest winds on the west side of trough; (B) strongest winds in southern portion of trough; (C) strongest winds on east side of trough; (D) excessive contour gradients.

Convergence and Divergence Above 500 Millibars

Study the 300-mb (or 200-mb) chart to determine areas of divergence and convergence and note these areas and the relative strength and extent of each for consideration with the movement of the long wave and progging of contours.

Convergence and divergence are covered in chapter 4 of this training manual. As a review of the effects of convergence and divergence on the changes in the intensity of troughs and ridges, we have the following rules. Refer to chapter 4 for illustrations of these rules.

1. Divergence and upper height falls are associated with high speed winds approaching cyclonically curved weak contour gradients. Divergence results in height falls to the left of the high speed current.
2. Convergence and upper height rises are associated with low speed winds approaching straight or cyclonically curved strong contour gradients and with high speed winds approaching anticyclonically curved weak contour gradients.

FORECASTING THE MOVEMENT OF PRESSURE SYSTEMS ALOFT

In this section we are concerned with the movement of closed pressure systems aloft. In this category we find both anticyclonic and cyclonic centers.

Movement of Highs

SEMIPERMANENT HIGHS.–The semipermanent subtropical highs are ordinarily not subject to much day to day movement. When a subtropical high begins to move, it will move with the speed and in the direction of the associated ridge. The movement of the ridge (long wave) has already been discussed. Too, seasonal movement, though slower and over a longer period of time, should be considered. These highs tend to move poleward and intensify in the summer and move equatorward and decrease in intensity in the winter.

BLOCKS.–Blocks will ordinarily persist in the same geographic location unless the closed contours of the high are strongly eccentric; that is, they are of uniform shape and a number of closed contours are present. Blocks will move in the direction of the strongest winds when contour spacing is asymmetric; their motion is eastward when the westerlies are strongest and westward when the easterlies are strongest. The speed of movement of these systems can usually be determined more readily and accurately by extrapolation. Extrapolation should be used in moving the highs under any circumstance, and the results of this extrapolation should be correlated with the other aspects of the movement and intensity prognoses of long and short waves.

Some indications of intensity changes which are exhibited by low tropospheric charts (700-500 mb) are as follows: intensification will occur with warm air advection on the west side; weakening and decay will occur with cold air advection on the west side; and there is little or no change in the intensity if the isotherms are symmetric with the contours. This low tropospheric advection is not the cause of the intensity change but is only a "sign." The cause is at higher levels; for example, intensification is caused by high-level cold advection and/or mass convergence. Under low index situations a blocking high will normally exist at a northern latitude and will have a pronounced effect on the systems in that area; in general it will slow the movement. Under high index situations there is a strong west to east component to the winds and systems will move fairly rapidly.

Movement of Closed Lows

The permanent Icelandic and Aleutian lows undergo little movement. These permanent lows will contract or expand in area; occasionally split into two separate low cells; become eccentric; or become elongated along the horizontal during a high zonal index type of situation. The north-south displacement is due primarily to seasonal effects. The movement of these permanent lows is derived primarily from extrapolation.

EXTRAPOLATION.–Extrapolation can be used at times to forecast both the movement and the changes in intensity of upper air closed lows. This method should be used in conjunction with other methods to arrive at the predicted position and intensity. Figure 8-6 shows some examples of simple extrapolation of both movement and intensity. Remember there

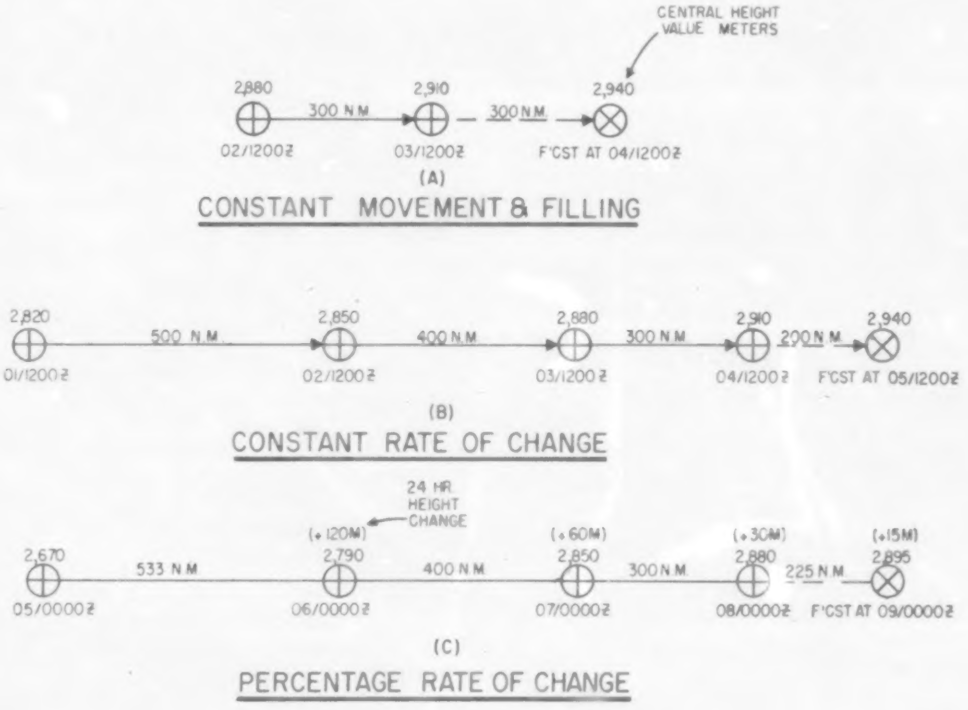
CENTRAL HEIGHT VALUE METERS
2,880
300 N.M.
2,910
300 N.M.
2,940
02/1200Z
03/1200Z
F'CST AT 04/1200Z
(A)
CONSTANT MOVEMENT & FILLING
2,820
500 N.M.
2,850
400 N.M.
2,880
300 N.M.
2,910
200 N.M.
2,940
01/1200Z
02/1200Z
03/1200Z
04/1200Z
F'CST AT 05/1200Z
(B)
CONSTANT RATE OF CHANGE
24 HR HEIGHT CHANGE
(+120M)
(+60M)
(+30M)
(+15M)
2,670
533 N.M.
2,790
400 N.M.
2,850
300 N.M.
2,880
225 N.M.
2,895
05/0000Z
06/0000Z
07/0000Z
08/0000Z
F'CST AT 09/0000Z
(C)
PERCENTAGE RATE OF CHANGE

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Figure 8-6.–Simple extrapolation of the movement and intensity of a closed low on the 700-mb chart. (A) Constant movement and filling; (B) constant rate of change; (C) percentage rate of change.

are many variations to these patterns and each case must be treated as an individual one.

Figure 8-6 (A) illustrates a forecast in which a low is assumed to be moving at a constant rate and filling at a constant rate. Since the low has moved 300 nautical miles in the past 24 hours, it is assumed that it will move 300 nautical miles in the next 24-hour period. Similarly, since the central height value has increased by 30 meters in the past 24 hours, you would forecast the same 30 meters increase for the next 24 hours. While this procedure is very simple, it is seldom sufficiently accurate. It is often refined by consulting a sequence of more than two charts to determine a rate of change.

This principle is illustrated in figure 8-6(B). By consulting the previous charts we find the low is filling at a rate of 30 meters per 24 hours and therefore this constant rate is predicted to continue for the next 24 hours. However, the rate of movement is decreasing at a constant rate of change of 100 nautical miles in 24 hours. Hence this constant rate of change of movement is then assumed to continue for the next 24 hours so the low is now predicted to move just 200 nautical miles in the next 24 hours.

Oftimes neither of these two situations exist and both the rate of change of movement and the height center change occur at a percentage rate. This is illustrated in figure 8-6(C). From a sequence of charts 24 hours apart it is shown that the low is filling at a decreasing rate and moving at a decreasing rate. The height change value is 50 percent of the value 24 hours previously on the successive charts and the rate of movement is 75 percent. We then assume this constant percentage rate to continue for the next 24 hours so the low is forecast to move 225 nautical miles and fill only 15 meters.

Accelerations may be handled in a similar manner as decelerations in figure 8-6. Also, a sequence of 12-hour charts could be used in lieu of 24-hour charts to determine past trends.

CRITICAL ECCENTRICITY.–When a migratory system is unusually intense, the system may extend vertically beyond the 300-mb level. Advection considerations, contour-isotherm relationships, convergence and divergence considerations, and the location of the jet max will yield the movement vector. These principles are applied in the same manner as they are when the movement of long waves are determined. The eccentricity formula may be applied to derive a movement vector, but only when a nearly straight eastward or westward movement is apparent. Migratory lows also follow the steering principle and the mean climatological tracks. The climatological tracks must be used cautiously for the obvious reasons. The rise and fall centers of the time differential charts are of great aid in determining an extrapolated movement vector, and extrapolation is the primary method by which the movement of a closed low is determined.

Certain cutoff lows and migratory dynamic cold lows lend themselves to movement calculation by the eccentricity formula. The conditions under which this formula may be applied are: The low must have one or more closed contours (nearly circular in shape), and the strongest winds must be directly north or south of the center. The location of the max winds determines the direction of movement. When the strongest winds are the easterlies north of the low, the low moves westward; when the strongest winds are the westerlies south of the low, the low will move eastward. The low will also move toward the weakest diverging cyclonic gradient and parallel to the strongest current. Systems moving eastward must have a greater speed in order to overcome convergence upstream–there is normally convergence east of a low system.

The eccentricity formula is written:

$$Ec = V - V' - 2C$$

or

$$2C = V - V' - Ec$$

where

Ec is the critical eccentricity value.
V is the wind speed south of the closed low.
V′ is the wind speed north of the closed low.
C is the speed of the closed low (in knots).

In order to obtain the value of C, it is necessary to determine the latitude of the center of the low and the spread (in degrees latitude) between the strongest winds in the low and the center of the low. Enter table 8-2 with these values; and apply the tabular value thus obtained to the critical eccentricity formula to obtain 2C, thus C. In determining the critical eccentricity of a system, it is necessary to interpolate both for latitude and the spread. A negative value for C indicates westward movement; a positive value, eastward movement.

Table 8-2.—Critical Eccentricity Value.

Latitude (degrees)	Spread (degrees latitude)				
	1°	3°	5°	10°	20°
80	.1	.9	2.5	—	—
70	.2	1.8	4.9	19.5	80.0
60	.3	2.6	7.1	27.0	115.0
50	.4	3.3	9.1	37.0	150.0
40	.4	4.0	10.9	43.5	175.0
30	.5	4.5	12.3	50.0	200.0
20	.5	4.9	13.3	53.0	—
10	.6	5.2	14.0	56.0	—

GRID METHOD.—The grid method as explained previously for short waves may also be used to prog the movement of closed lows. However, the equations must be modified. The same "X" factors are used for both methods. The procedure is as follows:

1. Place the center of the grid over the center of the low with the line DOC lined up directly north-south.

2. Heights are then read off the 500-mb height field at all points A through H.

3. Using the following formulas, substitute heights for the letters in each equation noting that plus values indicate east/south movement and minus values west/north movement.

(a) "X" [(A − B) + (C − D) + (E − F)] = east or west movement in degrees latitude.

(b) "X" [(B − F) + (G − H) + (A − E)] = north or south movement in degrees latitude.

4. The same correction of minus one degree latitude per 24-hour period should be applied to the east-west component only when an eastward movement is indicated.

5. The following sources of error have been noted:

a. Lows centered in troughs flanked by an intense ridge.

b. Small lows which move southeastward around a well developed ridge. These lows will move more rapidly than forecast in every case.

c. Lows centered in long-wave troughs between strong ridges.

d. Cases in which rapid and intense sea level development occurred just ahead of the low.

LOCATION OF THE JETSTREAM.—As long as a jet maximum is situated or moves on the western side of a low this low will not move out. When the jet center has rounded the southern periphery of the low and is not followed by another center upstream, the low will move out rapidly and fill.

ISOTHERM-CONTOUR RELATIONSHIP.—Little movement will occur it the isotherms and contours are symmetrical (no advection). The lows will strengthen and retrogress if cold air advection occurs to the west and progress eastward and fill if warm air advection occurs to the west.

FORECASTING THE INTENSITY OF PRESSURE SYSTEMS ALOFT

Many of the same considerations as explained in the previous section for the movement of closed centers aloft also apply to progging their intensity, as in most cases intensity and movement are considered simultaneously. Extrapolation and the use of time differentials aid in forecasting the change and the magnitude of falls in moving lows. Again, the rise and fall indications must be correlated with advection considerations, divergence indications, and the other factors discussed previously.

Progging the Intensity of Pressure Systems

HIGHS.—Highs undergo little or no change in intensity when isotherms are symmetric with contours.

Highs intensify when warm air is advected on the west side of the high.

Highs break down and weaken when cold air is advected on the west side.

Blocking highs usually intensify during westward movement and weaken during eastward movement.

Convergence and height rises downstream in the troughing area occur when high-speed winds with a strong gradient approach low-speed winds in an anticyclonic weak gradient. This is often the case in ridges, where the west side of the ridge harbors the high-speed winds, and the ridge intensifies as a result of this situation due to the accumulation of mass. Accumulation of mass due to convergence occurring with this situation has also been termed "overshooting." This high-level anticyclogenesis can be detected at the 500-mb level, but the 300-mb level is better suited for this and will yield a more reliable result, for it is the addition or removal of mass at higher levels which determines the height of the 500-mb contours.

The rise and fall centers of the time differential chart indicate the changes in intensity for extrapolation, both sign and magnitude. If other indicators agree with the rise and fall indications, progging the intensity (or the change in intensity of pressure systems aloft) is most easily and accurately accomplished by extrapolating from the time differential. The magnitude of the height rises can be adjusted when other indications reveal that a slowing down or a speeding up of the anticyclogenetic processes is occurring and expected to continue.

LOWS.—Lows and cutoff lows intensify when COLD air is advected to the west and they weaken (or fill) when WARM air is advected to the west.

Lows weaken when a jet maximum rounds the southern periphery of the low and when the jet max is on the east side of the low, if another jet max does not follow.

Lows strengthen when the jet max remains on the west side of the low. The qualification here is that the jet max to the west of the low may not be preceded by another on the southern periphery or eastern periphery of the low, for this indicates no change in intensity.

The 24-hour rise and fall centers aid in extrapolating both the change and the magnitude of falls in moving lows. Again, the rise and fall indications must be correlated with advection considerations, divergence indications, and the indications of the contour-isotherm relationships.

Progging the Formation of Pressure Systems

HIGHS.—Anticyclogenesis is ordinarily not a problem in progging the 500-mb level (and higher levels) except for the formation of blocking highs.

The shallow cold air masses of polar and Arctic origin generally give no indication of formation at the 500-mb level, and higher, and usually do not extend to this level either. The Aerographer's Mate must look for indications of this type of anticyclone at the lower levels.

High level anticyclogenesis is indicated when low-level warm advection is accompanied by stratospheric cold advection, and this situation has primary application to the formation of blocks, for high-level anticyclogenesis is primarily associated with the formation of blocks and the intensification of the ridges of the subtropical highs. The intensification of the subtropical highs (the ridges) has already been treated.

Blocks should normally be forecast to form only over the eastern part of the oceans in the middle and high latitudes. There should be present north to northwestward warm advection and northward moving height rise areas or persistent height rises at the higher latitudes, especially when southerly or southeasterly jets are present upstream.

The shallow anticyclones of polar or Arctic origin give indications of their genesis primarily on the surface and the 850-mb charts. The area of genesis will show progressively colder temperature at the surface and aloft; however, the drop in the 850-mb temperatures does not occur at the same rate as at the surface, an indication that a very strong inversion is in the process of forming. The air in the source region must be relatively stagnant.

LOWS.—Cyclogenesis has a number of indicators, and the greater the number of indicators in agreement, the greater the chance of success in progging cyclogenesis. They are as follows:

1. Areas of divergence exist at higher levels.

2. Jet maximums on the west side of a low indicate deepening and southward movement.

3. Cold advection in the lower troposphere and warming in the lower stratosphere are associated with the formation of or intensification of lows.

The remaining problem in forecasting the formation of lows is the problem of progging the formation of new cutoff lows. The indications which call for progging new cutoff lows are:

1. They generally form only off the southwestern coast of the United States and the northwestern coastal area of Africa.

2. The upstream ridge intensifies greatly and assumes an "overlapping" orientation. An intensifying ridge upstream is recognized when that ridge contains strong, strengthening, and sustained southwesterly flow.

3. Strong northerlies are situated in the west side of the trough.

4. Height falls move south or southeastward.

5. Strong cold advection appears on the west side of the upper trough.

Constructing Prognostic Contours

The constant pressure prognostic chart is about to take form. The steps leading to this, the final step, have been discussed. The prognostic position of the long wave troughs and ridges were determined and plotted on the tentative prognostic chart. The position of the highs, lows, and cutoff centers were then determined and plotted on the tentative prognostic chart. Short waves were treated in a similar fashion. Contours are constructed next, connecting these systems in the fashion exhibited by the constant pressure charts. The pattern of the contours is largely determined by the position of the long waves, short waves, and closed pressure systems. The height values of the contours are determined by actual changes in intensity of the system and by simple advection of higher or lower values. Contours are drawn in accordance with the following six steps:

1. Outline the areas of warm and cold advection in the stratum between 500 and 200 mb and move the thickness lines at approximately 50 percent of the indicated thickness gradient in the direction of the thermal wind.

2. Tentatively note at several points on the prognostic chart the increase or decrease in the height of the constant pressure surface over existing height values.

3. Move the areas of 24-hour height rises and falls at the speed of the short waves and note at several key points the amount and direction of the height change from the current chart.

4. Adjust the advected height changes to the height changes indicated by the rise and fall areas and adjust these in turn for position to the position of long waves, pressure systems, and short waves.

5. Heights for selected points at 500 mb are then extrapolated on the basis of the 24-hour time differential indications and advection considerations, provided that they are justified by the indications of high-level convergence and divergence. When the contributions from advection and time differentials are not in agreement with convergence and divergence (which is rarely the case), adjust the contribution of each and use this adjusted value.

6. Finally, adjust the height values thus indicated to the forecast intensity of the systems. These adjustments can lead to one of three possibilities for each system:

a. All factors point toward intensification (deepening of lows-filling of highs).

b. One factor washes away the contribution made by another and the system remains at or near its present state of intensity.

c. All factors point toward weakening of the system.

7. Sketch the preliminary contours, connecting the forecast positions of the long waves, short waves, and the pressure systems with contours with the height values determined by steps 1 through 5 above.

The last step in the construction of a constant pressure prognostic chart is to check the chart for these following points: The chart should follow continuity from the existing pattern. It should be vertically consistent and rational in the horizontal. It should not deviate from the seasonal pattern unless substantiated beyond doubt; and unless indicators dictate otherwise, it should follow the normal patterns.

Now draw the smooth contours, troughs, ridges, highs, and lows; and adjust the gradients.

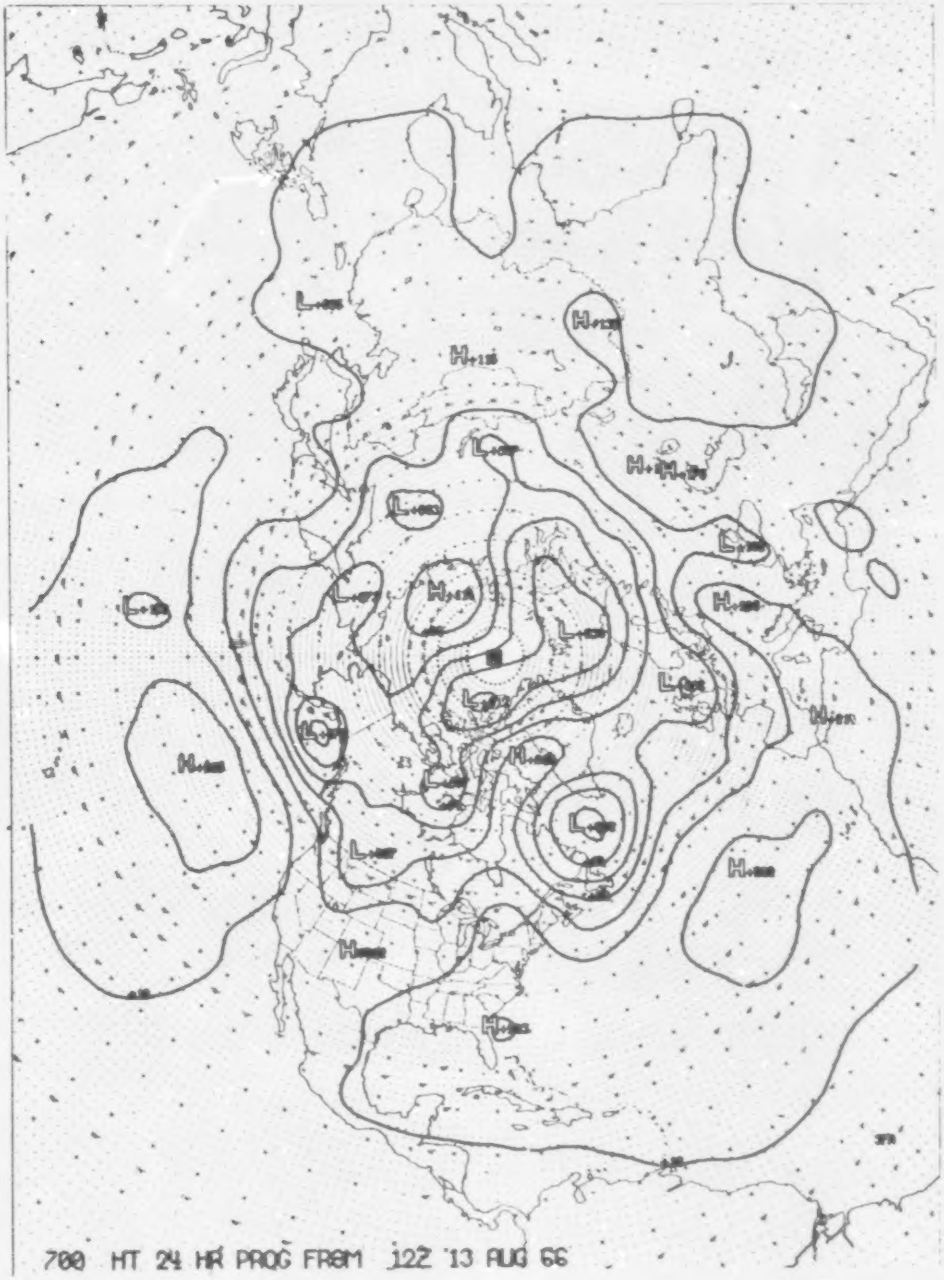
700 HT 24 HR PROG FROM 12Z 13 AUG 66

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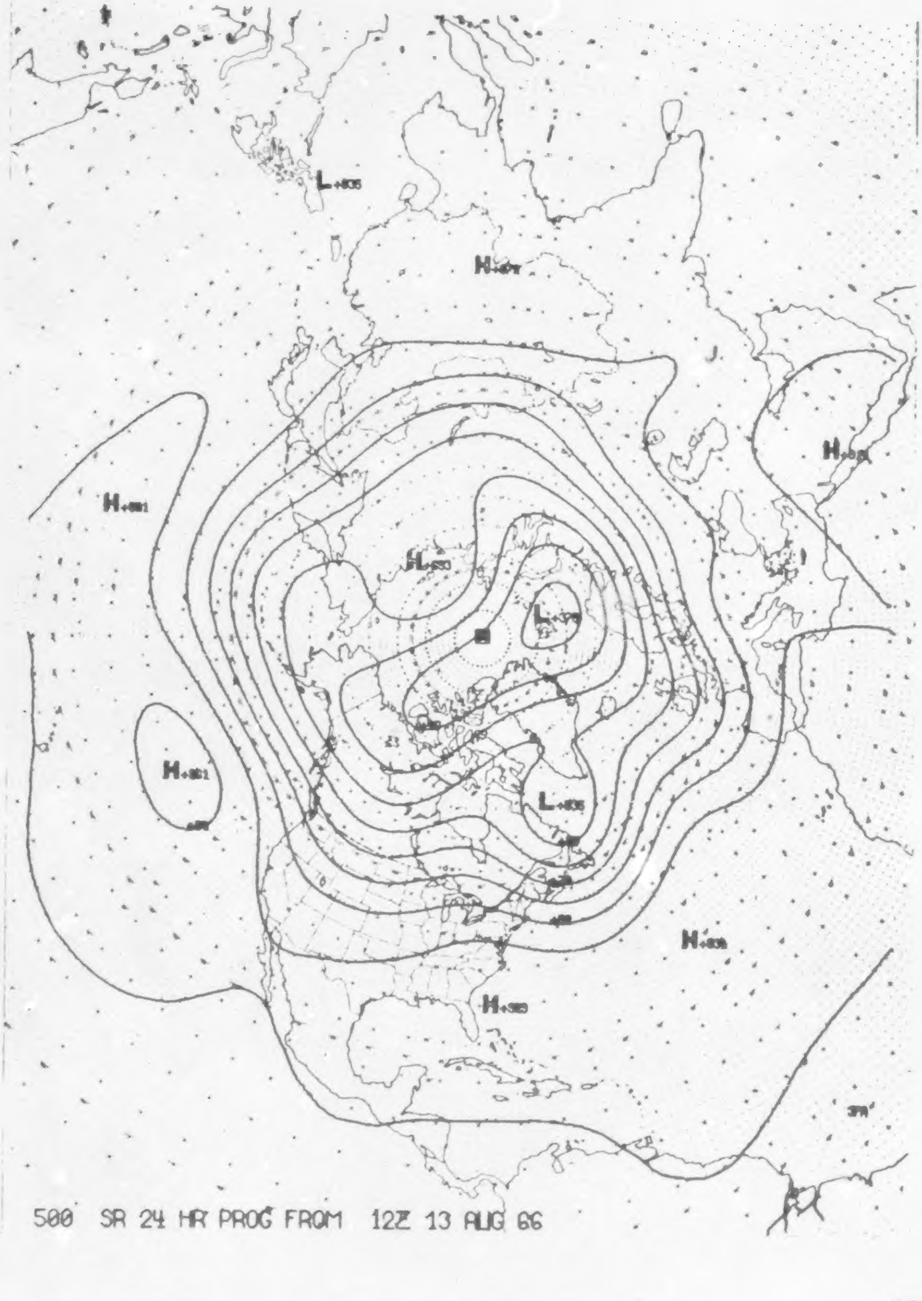
Figure 8-7.—700-mb height 24 hour prognosis. Contour interval 60 meters.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500 SR 24 HR PROG FROM 12Z 13 AUG 66

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Figure 8-8.—500-mb residual or long wave 24 hour prognosis. Contour interval 60 meters.

Application of Satellite Products

Satellite cloud pictures provide the forecaster with information that may be utilized along with already discussed techniques in forecasting the movement and intensity of troughs, ridges, and pressure systems aloft. As discussed in previous chapters the satellite pictures should be compared with the analyzed products so that these charts reflect a true picture of the atmosphere. Corrected charts utilized in preparing prognostic charts and forecasts will insure a greater degree of accuracy and validity.

Some of the more significant features which can be useful in assisting the forecaster in producing his prognostic upper air charts are positive vorticity advection maximum (PVA maximum), the cloud patterns associated with the upper level troughs and ridges, as well as cloud patterns that are indicative of the wind flow aloft.

All of these features and how to utilize them are discussed in Air Weather Service Technical Report 212, Application of Meteorological Satellite Data in Analysis and Forecasting, as well as Chapter 7 of this manual. Naval Air Systems Command, Naval Weapons Engineering Support Activity (FAMOS) publication, Guide for Observing the Environment with Satellite Infrared Imagery, (NWRF-F-0970-158), as well as Tech Report 212, contains series of satellite photographs and associated charts for various phenomena with detailed explanations.

Forecasters should become familiar with information contained in both of these publications to apply the satellite photographs most-effectively to their prognostic charts.

Application of Computer Products

The National Meteorological Center produces and distributes via the National Facsimile Network a number of varied computer prognostic charts. Some of the more useful upper air prognostic charts are the 24, 36, 48, and 72 hour constant pressure prognostic charts.

Various level upper air prognostic charts with varying valid times are transmitted via facsimile daily. Items included on the charts will vary on an individual basis, with respect to the following: contours for the particular height, isotachs, and isotherms.

The forecaster may utilize these charts directly for preparing his forecast or in conjunction with his own prepared products. A complete listing of charts available with description is found in the National Weather Service Forecaster's Handbook No. 1-Facsimile products.

The Fleet Numerical Weather Central also prepares a large number of computer products for upper air forecasting. These include the various constant pressure level progs, 500-mb small scale disturbance progs (SD), and 500-mb residual or long wave prognosis (SR). The U.S. Naval Weather Service Computer Products Manual (NAVAIR 50-1G-522) contains a detailed listing of charts available. Figures 8-7 and 8-8 are examples of the computer products produced by the Navy.

SINGLE STATION FORECASTING

With the advanced methods of communications and weather observing techniques most stations will have an abundance of information for preparation of forecasts. However, aboard ship single station forecasting techniques are frequently employed due to sparsity of reports or communications problems.

Close analysis of upper air soundings and surface conditions will provide the forecaster with a means of determining changes that are occurring in the atmosphere and utilize this information in preparation of his forecasts.

Detailed information on single station forecasting is contained in the Handbook of Single Station Forecasting Techniques (NA 50-1P-529).

CHAPTER 9

FORECASTING SURFACE SYSTEMS

With the upper air prognosis completed, the next step is to construct the surface prognostic chart. Since there are more data available for the surface chart, and this chart is chiefly the one which the Aerographer's Mate will base his forecast, prognosis of this chart should be carefully constructed to give the most accurate picture possible for the ensuing period. The surface prog may be constructed for periods up to 72 hours, but normally the period is 36 hours or less. In local terminal forecasting, the period may range from 1 to 6 hours. Time ranges over 36 hours are highly subjective and are of little value because of their doubtful reliability.

Construction of the surface prog consists of three main tasks. They are:

1. Progging the formation, dissipation, movement, and intensity of pressure systems.
2. Progging the formation, dissipation, movement, and intensity of fronts.
3. Progging the pressure pattern; that is, the isobaric configuration and gradient.

From an accurate forecast of the foregoing features, you should be able to forecast the weather phenomena to be expected over the area of interest for the forecast period.

FORECASTING THE FORMATION OF NEW PRESSURE SYSTEMS

The central problem of surface prognosis is to predict the formation of new low-pressure centers. This problem is so interrelated to the deepening of lows, that both problems are considered simultaneously when and where applicable. This problem mainly evolves into two categories. One is the distribution of fronts and air masses in the low troposphere, and the other is the velocity distribution in the middle and high troposphere. The rules applicable to these two conditions are discussed when and where appropriate.

For the principal indications of cyclogenesis, frontogenesis, and windflow at upper levels, refer to chapters 4 and 5 of this training manual.

The utilization of hand drawn analysis and prognostic charts for forecasting the development of new pressure systems is in many cases too time consuming to be effective. For most cases the forecaster will generally rely on satellite photographs and/or computer drawn prognostic charts to use in preparing this type of forecast.

In order to most effectively use satellite photographs it is necessary that the forecaster become familiar with satellite pictures, be able to interpret the pictures properly, and associate them with the corresponding surface phenomena.

Air Weather Service Technical Report 212 and the Navy Weather Research facility Detachment Suitland, Project FAMOS publication, Guide for Observing the Environment with Satellite Infrared Imagery (NWRF F 0970-158), both contain much useful information that the forecaster should be aware of.

The widespread cloud patterns produced by cyclonic disturbances represent the combined effect of active condensation from upward vertical motion and horizontal advection of a cloud. Storm dynamics restrict the production of cloud to those areas within a storm where extensive upward vertical motion or active convection is taking place. For disturbances in their early stages, the limits of the vertical motion distribution comprise the most important factor controlling cloud distribution. The comma shaped cloud formation which precedes an upper tropospheric vorticity maximum is an example. Here, the clouds are closely related to the upward motion produced by positive vorticity advection (PVA). In many cases this cloud may be referred to as the PVA Max. See figure 9-1.

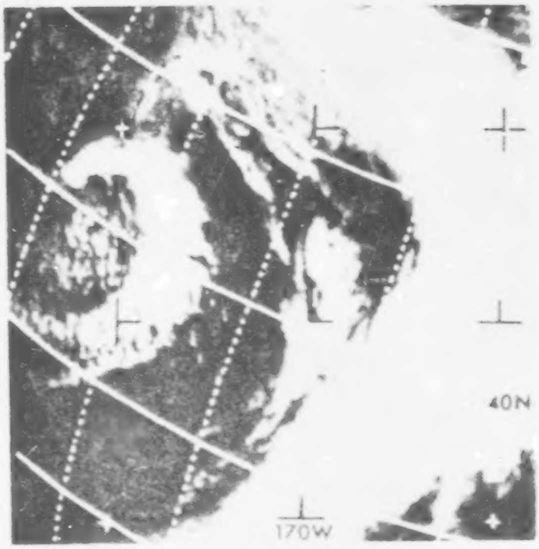

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Figure 9-1.—A well developed comma-shaped cloud is the result of a moving vorticity center to the rear of the polar front. The comma cloud is composed of middle and high clouds over the lower-level cumulus and is preceded by a clear slot.

Wave or low development along an already existing front may be detected from satellite photographs by correctly interpreting them. In figure 9-2 a secondary vorticity center is shown approaching a frontal zone.

The large cloud mass at A (fig. 9-2) is associated with a secondary vorticity center which has moved close to the front. Interaction between this vorticity center and the front will result in the development of another wave near B.

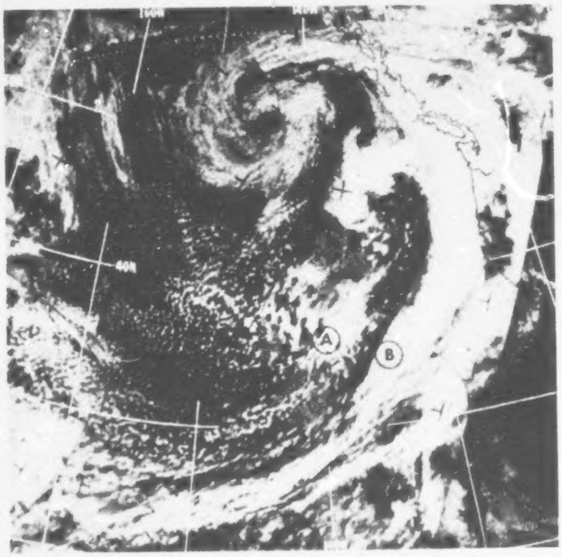

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Figure 9-2.—Frontal wave development.

By recognizing the vorticity center and determining its movement from successive satellite photographs the forecaster should be able to predict accurately and with confidence the formation of the second low pressure system.

Computer drawn charts also provide the forecaster with another tool for forecasting the development of new pressure systems. These prognostic charts may be used directly to prepare his forecast or he may utilize a number of them to construct other charts that will aid in forecasting.

On of the most useful charts in determining changes in surface pressure, frontogenesis, frontolysis, and development of new pressure systems is the advection chart. The normal methods of construction are time consuming; however, by utilizing computer charts the chart may be constructed in a fraction of the normal time.

The 700-mb and 1000- 500-mb thickness charts should be utilized in construction of the advection chart. The 700-mb contours have been found to approximate the mean wind vector between the 1000- and 500-mb level. On the

1000- 500-mb (H5-10) thickness chart the contours depict the following: thermal wind which blows parallel and downwind to the thickness lines, mean virtual temperature, and mean density. For advection purposes only mean virtual temperature should be used. Meteorologically speaking, we know that lines of greater thickness represent relatively warmer air than lines of less thickness. If the established advection pattern is replacing higher thickness values with lower thickness values, then it must be advecting cooler air (convergence and divergence not considered). The opposite of this is also true. The changing of thickness values can be determined by the mean wind vector within the layer of air. The 700-mb contours will be used as the mean wind vectors.

The advection chart should be constructed in the following manner:

1. Place the thickness chart over the 700-mb chart and line up properly.
2. Remembering that the 700-mb contours represent the mean wind vector, place a red dot indicating warm air at all intersections where the mean wind vector is blowing from higher to lower thickness values.
3. Using the same procedure place a blue dot at all intersections where the mean wind vector is blowing from lower to higher thickness values.

The placing of red and blue dots need not be done for the entire chart but only for the area of interest.

Now to utilize this advection chart it should be compared against the chart from the preceding 12 hours. From comparison of amount of red and blue dots it can be determined if there has been an increase or decrease in the amount of warm or cold air advection in a particular area as well as if there has been any change in the intensity of advection.

The advection type and amount as well as change then can be applied to determine the possibility of new pressure system development.

FORECASTING THE MOVEMENT OF SURFACE SYSTEMS

Whether you move the high- or low-pressure area first is a matter of choice for the forecaster. Most forecasters prefer to move the low-pressure area first and then the high-pressure area.

MOVEMENT OF LOW-PRESSURE AREAS

Lows determine, to a large extent, the frontal positions. They also determine a portion of the isobaric configuration in highs by virtue of the fact that gradients readjust between the two and the high seems to give way to the low. Although highs and lows are each possessed of their internal energy for motion, there is present a source of energy, derived from their interaction and which might be described as derived from a rotating earth; that is, independent of the systems. As a result of knowing the interplay of energy in the systems, meteorologists were able to evolve rules and methods for progging the movement, formation, intensification, and dissipation of lows.

Extrapolation

Extrapolation is also called the path method. First and foremost in your forecast of the movement of the low should be its past history. It is a record of movement of the pressure centers and attendant fronts, their direction and speed, and their intensifications. From this past history we can draw many valid conclusions as to the future behavior of the system and its future motion. This technique is valid for both highs and lows for short periods of time.

The general procedure for the extrapolation of low-pressure areas is outlined below. Although only movement is covered, the central pressures with anticipated trends could be added to obtain an intensity forecast.

1. Trace in at least four consecutive positions of the center at equal time intervals (6- or 12-hr intervals). Place an encircled X over each one of these positions and connect them with a dashed line. (See fig. 9-3.) Knowing the speed and direction, as obtained from past charts, it can easily be deducted where the position of highs or lows are on the present chart and extrapolated movement for the ensuing period can be made on this basis. One word of caution—straight linear extrapolation is seldom valid

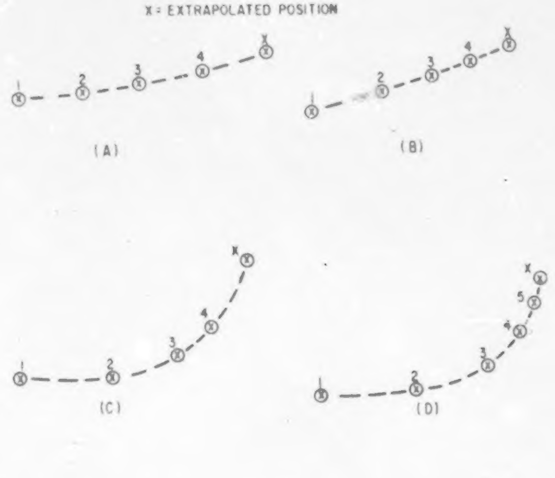

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Figure 9-3.–Example of extrapolation procedure. X is the extrapolated position.

beyond 12 hours. Beyond this, modifications, intensification, deceleration or acceleration, and changes in the path must be taken into consideration. It is extremely important that valid history be followed from map to map. Systems do not normally appear out of nowhere, nor do they just disappear.

In short, the extrapolation procedure involves two basic operations: First, a free hand extrapolation of the path; and second, an adjustment based on a comparison between the present chart and the preceding one. For example, the prolonged path of a cyclone center must not run into a stationary or quasi-stationary anticyclone, notably the stationary anticyclones, over continents in winter. When the projected path points toward such anticyclones, it will usually be found that the speed of the cyclone center decreases and the path curves northward. This path will continue northward until it becomes parallel to the isobars around the quasi-stationary high. The speed of the center will be least where the curvature of the path is greatest. When the center resumes a more or less straight path, the speed again increases.

Isallobaric Indications

It has been a well-known fact for many years that lows tend to move toward the center of the largest 3-hr pressure falls. This is usually the point where the maximum warm air advection is taking place. One such method using the isallobaric maximums and minimums is discussed later in the chapter.

Figure 9-4 shows the movement of an occluding wave cyclone through its stages of development in relation to the surface pressure tendencies. This one factor cannot be used alone. Several methods including this indication should be utilized to arrive at the final forecast. Too, you should remember that the process depicted in this illustration takes place over several days, and many other factors enter into the subsequent movement.

Circular, or nearly circular, cyclonic centers generally move in the direction of the isallobaric gradient; that is, in the direction of the greater pressure falls. Anticyclone centers move in the opposite direction. The speed is directly proportional to the isallobaric gradient.

Troughs move in the direction of the isallobaric gradient, and ridges move in the opposite direction. (See fig 9-5)

Relative to Warm Sector Isobars

Warm, unoccluded lows move in the direction of the warm sector isobars if those isobars are straight. These lows usually have straight paths (fig. 9-6(A)), whereas old occluded cyclones usually have paths that are curved northward (fig. 9-6(B)). The speed of the cyclones approximates the speed of the warm air.

Whenever either of these rules is in conflict with upper air rules, it is better to use the upper air rules.

Relative to Frontal Movement

The movement of the pressure systems must be reconciled with the movement of the associated fronts if the fronts are progged independently of the pressure systems. Two general rules are in use regarding the relationship of the movement of lows to the movement of the associated fronts: First, warm core lows are steered along the front if the front is stationary or nearly so; and second, lows tend to move with approximately the warm front speed and somewhat slower than the cold front speed.

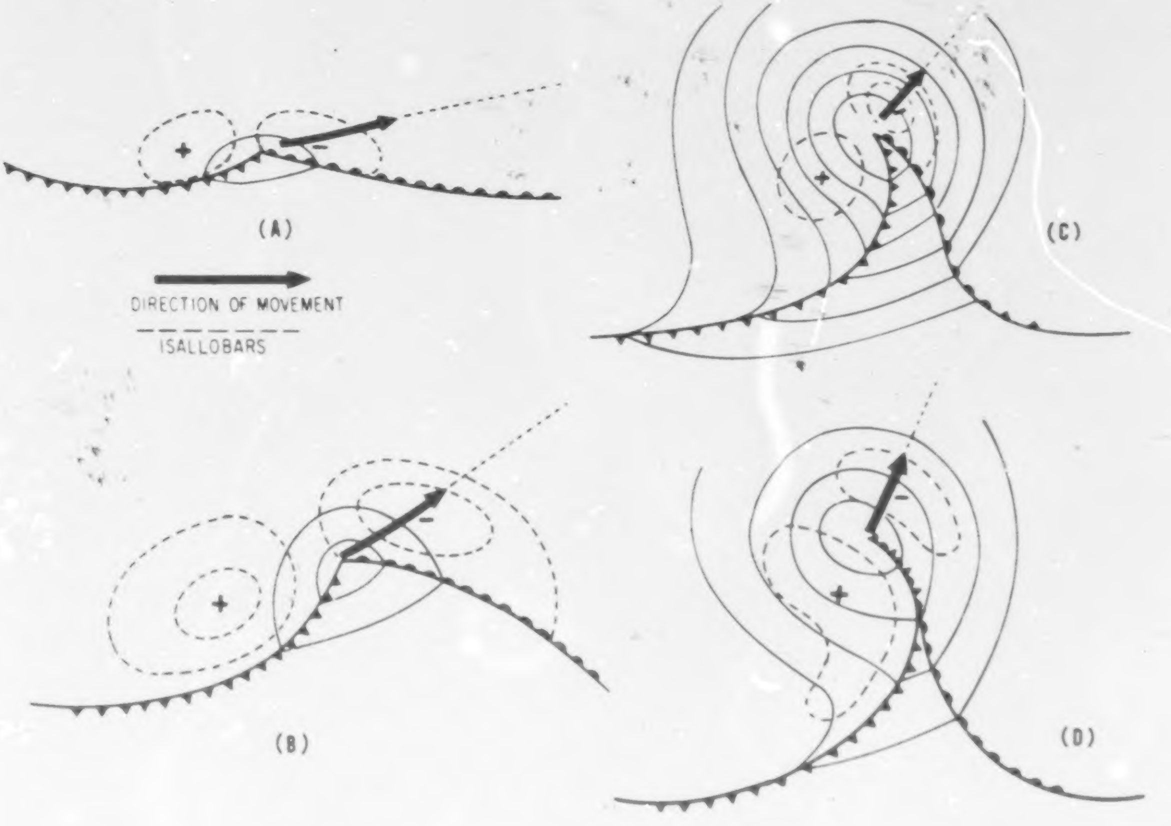

Figure 9-4.—Movement of occluding wave cyclone in relation to isallobaric centers.

Steering

Almost since the inception of synoptic weather maps, meteorologists have experimented with the idea of moving pressure systems with an upper-level steering current. This principle is based on the concept that pressure systems are moved by the external forces operating on them, just like a block of wood drifting in water. Thus, it might be said that a surface pressure system tends to be steered by the isotherms, contour lines, or streamlines aloft, by the warm sector isobars, or by the orientation of a warm front. This principle is nearly always applied to the relationship between the velocity of a cyclone and the velocity of the basic flow in which it is embedded.

The method works best when the flow pattern changes very slowly or not at all. If the upper flow pattern is expected to change greatly during the forecast period, we must first forecast the change in this pattern prior to forecasting the movement of the surface pressure systems. Do not attempt to steer a surface system by the flow of an upper level that has closed contours above the surface system. When using the steering method, we must first consider the systems which are expected to have little or no movement; namely, warm highs and cold lows. Then, consider movement of migratory highs

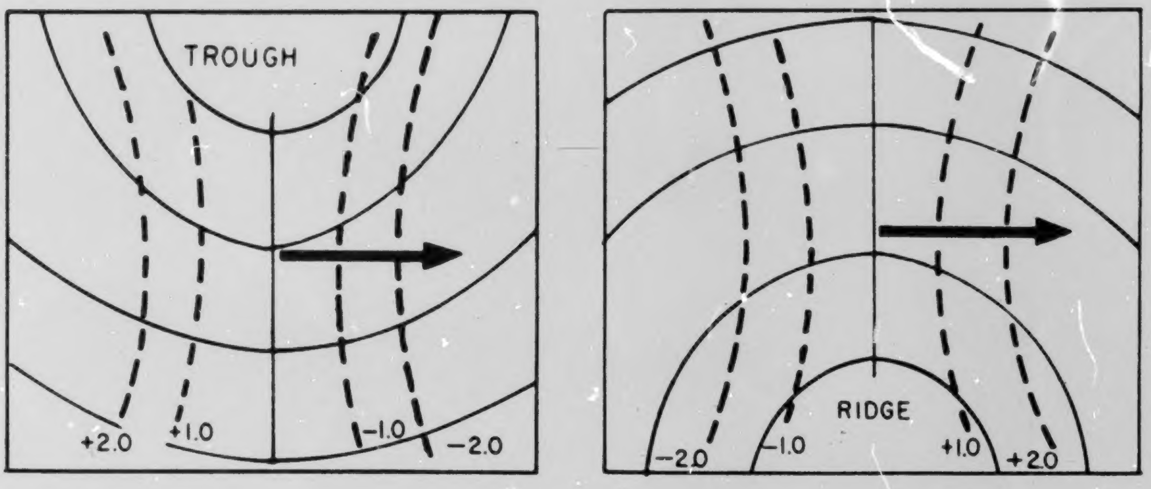

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Figure 9-5.—Movement of troughs and ridges in relation to the isallobaric gradient.

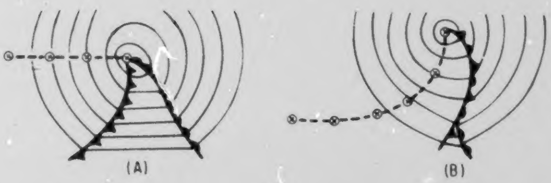

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Figure 9-6.—Movement of lows in relation to warm sector isobars. (A) Movement of warm sector lows; (B) movement of old occluded cyclones.

and lows; and finally, consider the changes in the intensity of the systems.

In studies using the steering technique, it was found in most cases that there was a displacement of the lows poleward and the highs equatorward of the steering current. Therefore, expect low-pressure centers, especially those of large dimension, to be deflected to the left and high-pressure areas deflected to the right of a westerly steering current. Over North America the angle of deflection averages about 15 degrees, although deviations range from 0 to 25 or even 30 degrees.

CAUTION: The steering technique should not be attempted unless the closed isallobaric minimum is followed by a closed isallobaric maximum some distance to the rear of the low.

The steering flow or current is the basic flow which exerts a strong influence upon the direction and speed of movement of disturbances embedded in it. The steering current or layer is a level or a combination of levels in the atmosphere which has a definite relationship to the velocity of movement of the embedded lower level circulations. The movement of surface systems by this flow is the most direct application of the steering technique. Normally, the level above the last closed isobar is selected. This could be the 700-, 500-, or 300-mb level. However, in practice it is better to integrate the steering principle over more than one level. The levels most often used are the 700- and 500-mb levels.

For practical usage, the present 700- or 500-mb chart should be used in conjunction with the 24- or 36-hr prognostic charts for these levels. In this way, changes both in space and time can be compensated for. For a direct application for a short period of time, transfer the position of the low center to the concurrent 700-mb chart. For direction, move the center in the direction of the contours downstream and slightly inclined to the left for low-pressure areas. Experience with moving systems of this type will soon tell you how much deviation should be made. For speed of the surface cyclone, average the basic current downstream over which the cyclone will pass (take into consideration changes in direction and speed of flow over the forecast period). Take 70 percent of this value for the mean speed for 24 hours. Move the low center along the contours as described above for this speed for 24 hours. This should be your position at that time.

For the 500-mb chart, follow the same procedure except use 50 percent of the wind speed for movement. If these two are not in agreement, take a mean of the two. There may be cases where the 500-mb chart is the only one used. In this case you will not be able to check the movement against the 700-mb chart.

The following empirical relationships and rules for steering should be taken into account when using the steering technique:

1. Warm, unoccluded lows are steered by the current at the level to which the closed low does not extend. When so steered, lows tend to move slightly to the left of the steering current.
2. Warm lows (unoccluded) are steered with the upper flow if a well-defined jet is over the surface center or if there is no appreciable fanning of the contours aloft. Low-pressure systems, especially when large, tend to move slightly to the left of the steering current.
3. Rises and falls follow downstream along the 500-mb contours, the speed is roughly half of the 500-mb gradient. Three-hour rises and falls seem to move in the direction of the 700-mb flow; while 24-hr rises and falls move with the 500-mb flow.
4. Cold lows, with nearly vertical axes, are steered with the upper low (in the direction of upper height falls), parallel to the strongest winds in the upper low, and toward the weakest contour gradient.
5. Occluded lows, the axes of which are not vertical, are steered partly in the direction of the warm air advection area.
6. A surface low which is becoming associated with a cyclone aloft will slow down, become more regular, and follow a strongly cyclonic trajectory.
7. Surface lows are steered by jet maximums above them and deviate to the left as they are so steered. They move at a slower rate than the jet maximum and are soon left behind as the jet progresses.
8. During periods of northwesterly flow at 700 mb from Western Canada to the Eastern United States, surface lows move rapidly from NW to SE bringing cold air outbreaks east of the Continental Divide.
9. If the upper height fall center (24-hr) is found in the direction in which the surface cyclone will move, the cyclone will move into the region or just west of it in 24 hours.

Direction of Mean Isotherms (Thickness Lines)

A number of rules have been formulated regarding the movement of low-pressure systems in relation to the mean isotherms or thickness lines. These rules are outlined as follows:

1. Unoccluded lows tend to move along the edge of the cold air mass associated with the frontal system which precedes the low; that is, it tends to move along the path of the concentrated thickness lines. When using this method, remember that the thickness lines will change position during the forecast period. If there is no concentration of thickness lines, this method cannot be used.
2. When the thickness gradient (thermal wind) and the mean windflow are equal, the low moves in a direction midway between the two. This rule is more reliable when both the thermal wind and the mean windflow are strong.
3. When the mean windflow gradient is stronger than the thickness gradient, the low will move more in the direction of the mean windflow.
4. When the thickness gradient is stronger than the mean windflow gradient the low will move more in the direction of the thickness lines.
5. With warm lows, the mean isotherms show the highest temperature directly over the surface low, which is about halfway between the 700-mb trough and ridge line. This means the mean isotherm and 700-mb isoheights are 90 degrees out of phase. Since warm lows move with the mean speed of the warm air above them, they will be rapidly moving systems.

6. On the other hand, if the highest mean temperatures occur under the 700-mb ridge (isotherms and contours in phase), the ridge itself is warm while the low is cold therefore a slowly moving low.

7. Lows move with a speed of approximately 50 percent of the thermal wind for the 1,000-500-mb stratum and approximately 75 percent of the thermal wind of the 1,000-700-mb stratum.

Movement of Lows in Relation to the Jetstream

Some of the rules for moving lows in relation to the jetstream position were mentioned previously under the section on steering. One basic rule however, states that highs and lows situated under the jetstream or very close to will mostly behave regularly and follow the steering calculation most closely. Minimum deviations occur when the upper flow does not change with time.

MOVEMENT OF HIGH PRESSURE AREAS

The literature available on the movement of high pressure areas is not so plentiful as that for low pressure areas. In general, the method for extrapolation of low pressure areas is applicable to the movement of high pressure areas as well.

The following are general considerations in forecasting the movement of high pressure systems.

A high or the portion of the high situated under a blocking high aloft remains very nearly stationary. High situated under a jetstream or very close to one are steered by the current aloft. Highs ordinarily do not follow the steering principle very closely. Cold shallow highs are steered more easily than the larger ones. The Canadian and Siberian highs move little when there is no jet max in their vicinity or above them, and they move rapidly when the jet max is present. Progressive warm highs move with a speed consistent with that of the major ridges aloft. Highs (and lows) are possessed of internal forces which assist in moving them. With straight westerly currents aloft, surface highs are displaced equatorward. Highs tend to move in the direction of and with the speed of the isallobaric centers; however, this rule is not very reliable because the isallobaric rises often follow the low rather than lead the high.

Steering is not used for high-pressure systems as widely as for lows because high-pressure cells do not have as great a vertical extent as low-pressure systems. However, steering seems to work about 75 percent of the time for cold highs.

FORECASTING MOVEMENT OF LOWS BY STATISTICAL TECHNIQUES

Since it requires many years of experience and a photographic memory to develop a mental catalog of weather patterns, a weather type or normal path classification is a boon to the inexperienced meteorologist as a means of quickly selecting analogous situations from the past for application to the present. There are many normals and average conditions to regulate behavior patterns of future movement and development. However, there are also many deviations from the north. The season of the year and topographical influences are factors to be considered. If we could catalog weather types or average types, and the storms obeyed these rules, it would greatly simplify the art of forecasting. However, this is just another tool in the integrated forecast. Use it, but do not lean too heavily on it.

Normal Tracks

In 1914, Bowie and Weightman published climatological tables of the average, by months, of the 24-hr speed and direction of cyclonic centers in the United States. The storms were classified with respect to the point of origin and the current location of the centers. Although these tables appear to be antiquated some of these types resemble relatively recent classifications and are therefore of some value to the present-day forecaster.

The Marine Climatic Atlases also contain average storm tracks for each month of the year for areas over the oceans of the world. Other publications are available which give average or normal tracks for other areas of the world.

Weather Types

This method of classifying significant features of the weather map with expected subsequent movements is also called the weather analogue method. The best known of these methods is the Elliott types.

In this typing system the broad features of the upper westerlies are recognized as the steering mechanisms of migratory cyclones and anticyclones. There now exists a set of 15 to 20 types for each of 5 zones extending from 135 degrees east through North America to 45 degrees east (Western Russia).

The North American Continent covers two types of zones. It is beyond the scope of this training manual to discuss this method in detail. However, this can be another valuable tool and aid in formulating a long range forecast.

FORECASTING MOVEMENT BY OBJECTIVE METHODS

Objective techniques are those techniques which employ certain predetermined parameters which are used in conjunction with graphs, tables, nomograms, or the like to arrive at a prediction. They should only be used and evaluated in the light of all of the other considerations. Objective methods do have several attractive features. Most of the methods available can be reduced to a simplified set of instructions with an attendant worksheet and give the forecaster a measure and degree of confidence not attained by using subjective methods alone. There are many objective methods available but most which are accessible to AG's appear to have been developed for the United States, with the geographical area east of the Rockies in particular. Only a few are covered here. Credit is given to the American Meteorological Society and to the authors of the article Evaluation of Techniques for Predicting the Displacement of Northeastward Moving Cyclones, BAMS, February 1956, by Wayne S. Herring and Wayne D. Mount, for permission to use information on the validity of some of these methods in general and for use of the Herring-Mount technique in particular in this section of this chapter. Some of the methods are covered only briefly as they are geographically restricted in application.

George's Method

J. J. George and Associates have developed a number of methods for forecasting the movement and intensity of certain categories of lows over the continental United States. These studies also contain information on the movement of highs. Information on these methods is contained in GRD Scientific Report No. 2, Contract No. AF19(122)-468, Further Studies on the Relationships Between Upper/Level Flow and Surface Meteorological Processes by R. J. Schafer and P. W. Funke. A simple explanation of the methods along with worksheets and diagrams is contained in each of these publications.

The method for moving lows yields a displacement forecast in terms of eastward and northward components of motion. These components are determined by graphical combination of (1) the zonal flow at 700 mb, (2) the meridional flow at 700 mb, (3) 12-hr changes in (1) and (2), and (4) a measure of the 700-mb temperature gradient, or the amplitude of the 700-mb trough. Objective application is comparatively easy and straightforward. The method gives a 30-hr forecast of motion.

Palmer Method

In 1948, W. C. Palmer developed a method for giving the 30-hr direction ray of cyclones based on the Bowie-Weightman tables. These data are for the winter months and for storms over the Eastern United States. The direction forecast is obtained from a graphical combination of the winter average 24-hr direction based upon the Bowie and Weightman tables, the past 6-hr direction of movement, and the orientation of a line joining the katallobaric and anallobaric centers associated with the cyclone. Forecast of 30-hr direction given by this combined analysis reduced the average error of 21.5 degrees, obtained from the original tables, to 16.1 degrees.

Herring-Mount Method

During the course of the evaluation on the movement of cyclone centers by the various methods, it was noted by the authors that cyclonic centers having a past 12-hr speed which was markedly slower or faster than normal speed, for the time of year, tend to approach the normal speed during the subsequent 30-hr period. Similarly it was noted that the storms of this class moved in a direction between that indicated by a line joining the isallobaric centers associated with the disturbance and the normal direction of movement 52 degrees east of north. (52 degrees east of north for Lambert Conformal charts and 50 degrees east of north for Polar Stereographic charts.)

A forecast method was defined simply by averaging the past 12-hr speed with the normal speed to give the 30-hr predicted speed, and the isallobaric direction was averaged with the normal direction of 50 or 52 degrees, depending on the chart used, to give the direction of forecast. The isallobaric direction is measured in degrees east of north from the meridian passing through the position of the storm on the surface chart. In this test, these results yielded the lowest direction, speed, and position errors of any of the rules tested. Figure 9-7 gives a simple illustration of the criteria by which this method can be applied, the table of average speeds, and an example of its use.

Prediction of Maritime Cyclones

This method is an empirically derived method for objectively predicting the 24-hr movement and change in intensity of maritime cyclones. The technique requires only measurement of the 500-mb height and temperature gradients above the current sea level center, and determination of the type of 500-mb flow within which the surface system is embedded. Full details of this method are described in The Prediction of Maritime Cyclones, NA 50-1P-545.

It is recommended that the deepening prediction be made first, as this will often give a good indication of movement.

The explosive intensification of maritime cyclones is a fairly common phenomenon, but is presently among the most difficult problems to forecast. Conversely, there are many situations in which it is important to predict the rapid filling of cyclones. This technique gives an objective method for predicting the 24-hr central change in pressure of those maritime cyclones whose initial positions lie north of 30 degrees north latitude. Further, the technique applies to the winter months only November through March, although it may be used with some degree of confidence in other months.

From the study, the following factors stood out as being the most important:

1. The location of the surface cyclone center with respect to the 500-mb pattern.
2. The strength of the 500-mb flow above the cyclone center.
3. The 500-mb temperature gradient to the northwest of the surface center.

Of the lows which intensified, the deepening was in general greater, the stronger the 500-mb contour and isotherm gradient. The study also indicated that the preferred location for filling cyclones is inside the closed 500-mb contours and that deepening cyclones favor the region under open contours in advance of the 500-mb trough. The remaining portions of the pattern indicate areas of relatively little change, except lows located under a 500-mb ridge line fill.

Recently Developed Techniques

Two techniques have been developed and published for the statistical prediction of cyclones and anticyclones over an area bounded by 30 and 60 degrees north and 60 and 110 degrees west. Both methods use a grid of finite intervals and certain statistical parameters at the surface and 500 mb to determine the future movement and intensity of these two systems. This is accomplished by a mathematical procedure involving stepwise multiple regression equations. Very promising results have been obtained by both methods.

The method for forecasting the movement and intensity of the surface cyclone centers in the area defined above was developed by K.W. Veigas and F.P. Ostby and published under the title Application of a Moving Coordinate Prediction Model to East Coast Cyclones, Journal of Applied Meteorology, Vol. 4, No. 2, pages 24-38. The statistical method of predicting the movement and changes in intensity of North American Anticyclones was based on the above method. It was developed at the U. S. Naval Post

HERRING & MOUNT SYSTEM FOR TYPE IV LOWS

1. Low must be located east of 95th meridian.
2. Low must have moved from southwest for past 12 hours.
3. Low must be under SW flow at 850, 700 and 500 mbs. (180° to 270°)

Measure number of degrees east of north of line between 3-hour anallobaric and 3-hour katallobaric centers. The average between this and 50 degrees east of north is 30-hour direction. (For Polar Stereographic Projection Chart). (52° for Lambert Conformal Charts).

The average between past 12-hour speed and normal speed for the time of year is average speed for next 30 hours.

****

NORMAL SPEEDS

NOV 1-15	17.5 kts	FEB 1-15	33.0 kts
NOV 16-31	23.5 kts	FEB 15-28	30.5 kts
DEC 1-15	28.0 kts	MAR 1-15	26.0 kts
DEC 16-31	31.5 kts	MAR 16-31	23.5 kts
JAN 1-15	34.0 kts		
JAN 16-31	35.0 kts		

* *

VISUAL AID

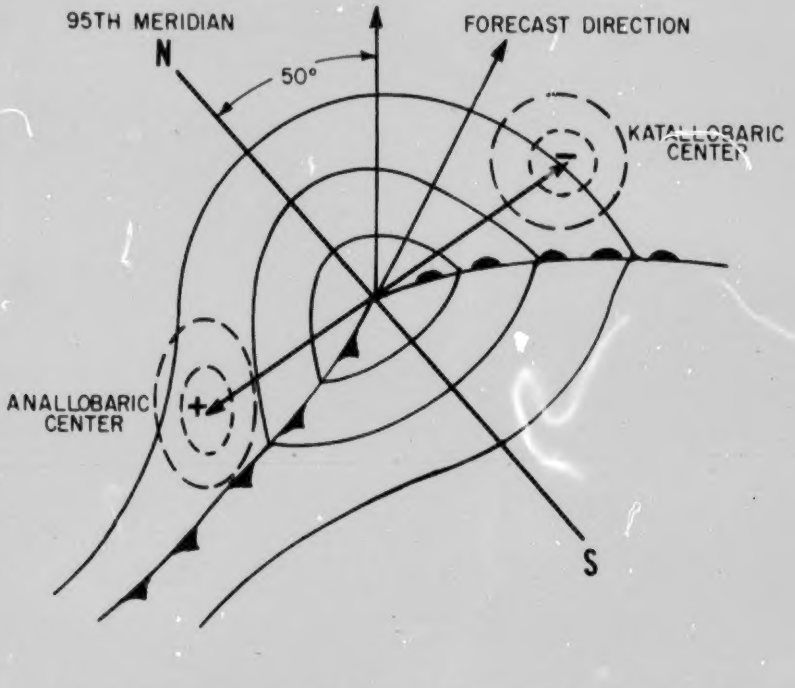

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Figure 9-7.—Illustration of the Herring-Mount technique for the moving of northeastward moving cyclones.

Graduate School, Monterey, Calif., and published under the title Statistical Prediction Methods for North American Anticyclones in the August 1963 issue of the Journal of Applied Meteorology, Vol. 2, No. 4, pages 508-516.

Although the use of calculators or computers would be desirable in working both methods, the grid points can be extracted and the values substituted into the equations and the computations made by simple arithmetic.

FORECASTING THE INTENSITY OF PRESSURE SYSTEMS

The changes in intensity of pressure systems at the surface are determined by what is occurring above the surface system, since the pressure measured at the surface is a measure of the weight of the atmosphere above.

EXTRAPOLATION

The pressure tendencies which are reported on the synoptic charts indicate the sum of the pressure change due to movement of the system plus that due to deepening and filling. If the exact amount of pressure change due to movement could be determined, it could be assumed that the system could continue to deepen or fill at the current rate of speed during the forecast period, then determining future pressure would be quite simple. However, it is not normally safe to assume that the current rate of change will continue nor just how much of the pressure change is due to movement. The pressure tendencies may be used in a qualitative sense for a quantitative approximation.

ISALLOBARIC INDICATIONS

Isallobaric indications from analyzed isallobars on the surface chart show the following relationships between the isallobars and the changes in the intensity of pressure systems:

1. When the 3-hr pressure falls extend to the rear of the low, the low is deepening.
2. When the 3-hr pressure rises extend ahead of the low, the low tends to fill.
3. When the pressure rises extend to the rear of a high-pressure cell or ridge line, the pressure system is increasing in intensity or the system is filling.
4. Since low-pressure systems usually move in a direction parallel to the isobars in the warm sector, and since the air mass in the warm sector is homogeneous it is possible to use the pressure tendency in the warm sector as an indication of the deepening or filling of the system. The effects of frontal passages must be removed. Therefore, if a low moves parallel to warm sector isobars, it can be said that the barometric tendency in the warm sector is equal to the deepening or filling. Also, the variation of the tendency in the warm sector, from the peak of the wave outward, can be used to indicate increasing or decreasing gradient.

It must be remembered that when using present tendency values for any of the rules, it is merely an indication of what has been happening and not what will necessarily take place in the future. Consequently, in using tendencies for indication of deepening or filling, it is necessary to study the past trend of the tendencies. The 12- and 24-hr pressure change charts are helpful in determining these values.

RELATIVE TO FRONTAL MOVEMENT

Wave cyclones form most readily on stationary or slow moving fronts. A preferred position is along a decelerating cold front in the region of greatest deceleration. Normally, the 700-mb winds are parallel to the front along this area.

Under conditions characteristic of the eastern Pacific, a secondary cyclone may develop with extreme rapidity. As a wave forms on the retarded portion of a cold front, the original well-marked warm front of the primary cyclone tends to fade out or become masked in the more or less parallel flow existing between the returning cold air from the high to the eastward and the original warm feeding current. In this stage of development the new secondary wave is actually a rapidly moving wave cyclone, but with no great deepening apparent as it moves along the front and few, if any, indications of occluding. As the new low moves eastward along the front, even though it is a shallow wave, the pressure gradients surrounding it will tighten and tend to sharpen the old masked warm front lying well to the east of the new wave itself. Later, as the wave moves rapidly eastward, it will pick up this resharpened warm front and the cold front of the wave cyclone in its field and then occlude on this resharpened warm front.

This occlusion exhibits well-marked phenomena such as rapid deepening as much as 10 to 15 mb in 12 hours and changing the structure of the original wave cyclone into one similar to that of the occluded cyclone. The resultant rapid deepening and increase in the cyclonic circulation results in a portion of the original polar front connection between the new and the old cyclone being destroyed. Figure 9-8 illustrates the process.

INDICATIONS ALOFT FOR DEEPENING AND FILLING OF SURFACE LOWS

Advective Pressure Changes

The role of temperature advection (that is, the transport of air of different density into a region) in contributing to the pressure change (or contour change) is somewhat controversial. On the one hand, low-level (usually 1,000-500 mb) warm advection is frequently cited as responsible for the surface pressure falls ahead of moving surface lows (the converse for cold advection); on the other hand, warm advection is frequently associated with rising contours of upper levels.

It is well known from the tendency equation that the pressure change at the SURFACE is equal to the pressure change at some UPPER LEVEL plus the change in mass of the column of air between the two. That is, if the pressure at some upper level remains UNCHANGED and the intervening column is replaced with warmer air, the mass of the whole atmospheric column (and consequently the surface pressure) decreases, and so does the height of the 1,000-mb surface.

As an example, assume that warm advection is indicated below the 500-mb surface (5,460 meters) above a certain station. If no change in mass is expected above this level, the pressure at 5,460 meters will remain unchanged, and so will the height of the 500-mb level. Suppose the 1,000- to 500-mb advection chart indicated that the 5,400-meter thickness line is now over the station in question and will be replaced by the 5,490-meter thickness line in a given time interval, that is, warm advection of 90 meters. The consequence is that the 1,000-mb surface, which is now 60 meters above sea level, will lower 90 meters to 30 meters below sea level and the surface pressure will decrease a corresponding amount, about 11 mb. Whenever the surface pressure is less than 1,000 mb, the 1,000-mb surface is below the ground and is entirely fictitious. In view of the above description of advective pressure changes, the following rules may be stated:

1. Warm advection between 1,000 and 500 mb brings falling sea level pressures, and the converse of this rule:
2. Cold advection between 1,000 and 500 mb brings rising sea level pressures.

Indications of Deepening From Vorticity

Cyclonic development and deepening are closely related to cyclonic flow aloft or to cyclonic vorticity aloft. If you recall from the discussion of vorticity in chapter 4 of this training manual, vorticity is the measure of the path of motion of a parcel plus the wind shear along the path of motion. Thus we have the following rules of the relationship of vorticity aloft to the deepening or filling of surface lows:

1. Surface pressure falls where advection of cyclonic relative vorticity takes place. Thus, it can be said that surface pressure falls where cyclonic vorticity decreases (is less) downstream.
2. Surface pressure rises where advection of more anticyclonic relative vorticity takes place. Thus, it may be said that surface pressure rises where cyclonic relative vorticity increases (is greater) downstream.
3. A wave will be unstable and deepen if the 700-mb wind field over it possesses cyclonic relative vorticity. A wave will be stable if the 700-mb wind over it possesses anticyclonic vorticity.
4. If there are several waves along a front, the one with the most intense cyclonic vorticity aloft will develop at the expense of the others.

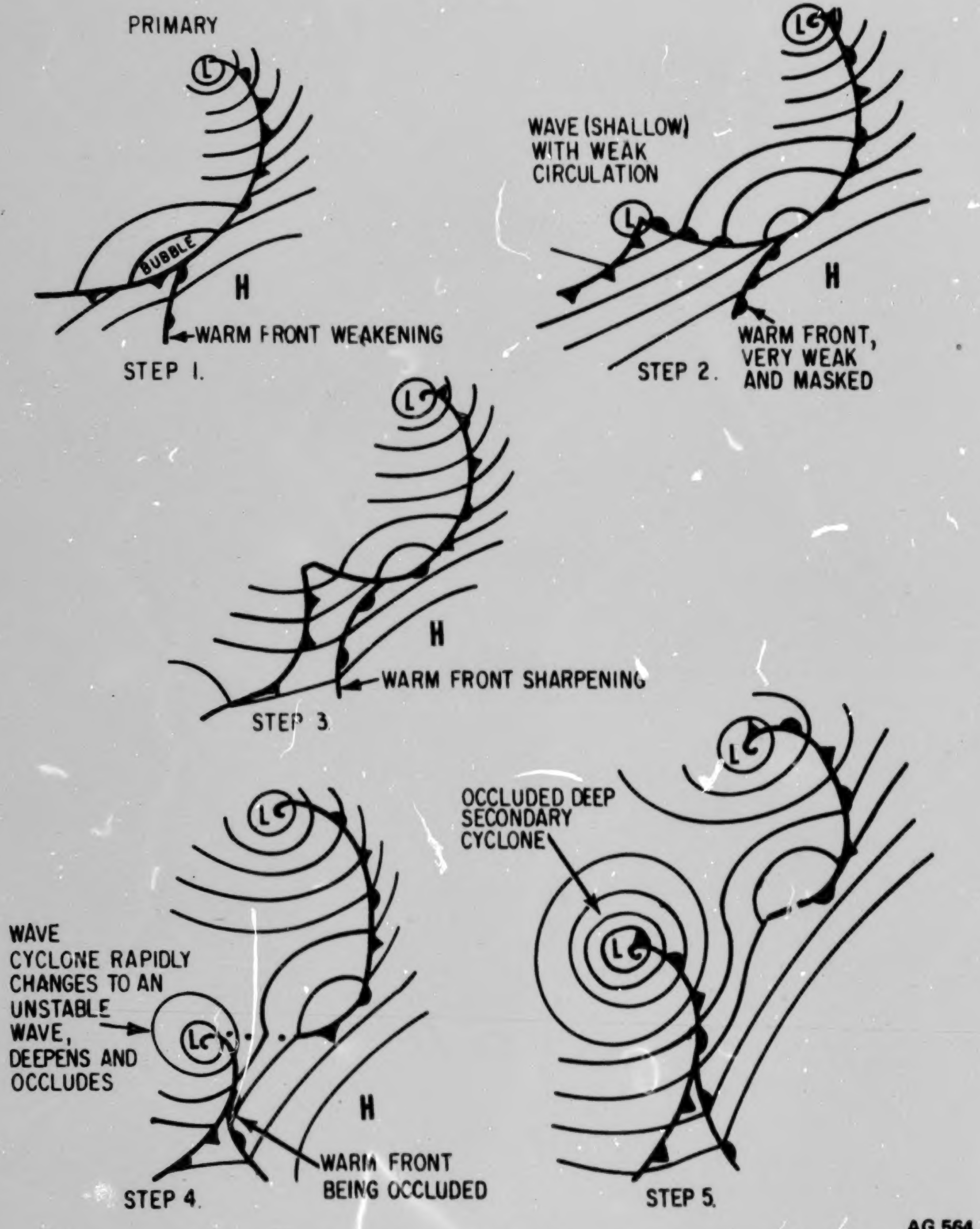
PRIMARY
L
BUBBLE
H
WARM FRONT WEAKENING
STEP 1.
L
WAVE (SHALLOW) WITH WEAK CIRCULATION
L
H
WARM FRONT, VERY WEAK AND MASKED
STEP 2.
L
H
WARM FRONT SHARPENING
STEP 3.
L
WAVE CYCLONE RAPIDLY CHANGES TO AN UNSTABLE WAVE, DEEPENS AND OCCLUDES
L
H
WARM FRONT BEING OCCLUDED
STEP 4.
L
OCCLUDED DEEP SECONDARY CYCLONE
L
STEP 5.
AG.564

Figure 9-8.—Illustration of secondary cyclone development over the eastern Pacific.

This is usually the one nearest the axis of the trough.

Deepening of Lows Moving Relative to Upper Contours

The amount of deepening of Eastern United States lows moving northeastward into the Maritime Provinces of Canada frequently can be predicted by estimating the number of contours at the 200- or 300-mb level that would be traversed by the surface low in the period of the prognostic chart. Low tropospheric cooling would compensate the stratospheric mass decrease about 40 percent. Therefore, the practice of taking about 60 percent of the height difference at 200 mb between current and prognostic positions proved successful in predicting the height change at the center of the surface low as it moved from its current to the prognostic location. For a close approximation, multiply the 200-mb current height difference in tens of meters by 3/4 and you have the surface deepening in millibars. For example: a 240-meter height difference at 200 mb results in a pressure change of 18 mb at the surface.

$$24\ 3/4 = 18 \text{ mb}$$

If FALLING heights aloft are indicated, the AMOUNT of fall does not have to be estimated, since deepening of the surface low will be greater than otherwise, and greater advective cooling, associated with occlusion, appears to compensate the upper falls. Therefore, the same method of approximation gives the magnitude of the deepening.

If RISING heights are indicated aloft over the expected low position, the AMOUNT MUST BE ESTIMATED in order to determine the ACTUAL height difference to which the rule will apply. In some cases the height rises aloft over the expected position of the low may be quite large, indicating the development of a high-latitude ridge aloft which tends to block the eastward progress of the low. This may result in rapid deceleration of the low, with filling and/or recurvature to the north. In such a case, the prognostic low position is revised in the light of the changing circulation aloft.

This technique works only when lows are expected to move northeastward out of a heat source, such as the Southern Plains. When a low moves into the Southern Plains from the west or northwest, there is frequently no compensatory cooling in the low troposphere, since the low is moving toward the heat source.

Some rules for filling and deepening of lows in relation to upper contours are stated in the following section:

1. Filling is indicated when a low moves into or ahead of the major ridge position of the 500-mb level.
2. Surface lows tend to fill when the associated upper-level trough weakens.
3. Lows tend to fill when moving toward values of higher thickness lines.
4. When the associated upper trough intensifies, deepening is indicated.
5. Lows deepen when moving toward lower thickness values.
6. Waves develop along fronts when the 700-mb windflow is parallel to the front or nearly so.
7. During periods of southerly flow at 700-mb, along the east coast of the United States, secondary storms frequently develop in the vicinity of Cape Hatteras.

High Tropospheric Divergence in Developing Lows

In the case of developing (dynamic) cyclones, horizontal divergence is a maximum in the 400- to 200-mb stratum, and the air above must sink and warm adiabatically in order to maintain continuity.

In the deepening of lows there must be removal of air at high levels due to divergence in the 400- to 200-mb stratum, which results in the stratospheric warming observed. Insufficient inflow at very high levels to compensate the subsidence results in the upper-level contour falls.

This is roughly the mechanism thought to be responsible for the development of low-pressure systems. The high-level decrease in mass overcompensates the low tropospheric increase in density; the high-level effect thus determines the reduction of pressure at the surface when lows are intensifying.

Stratospheric and Upper Tropospheric Decrease in Mass

The chief cause of deepening lows is the decrease in mass in the upper troposphere and the lower stratosphere. In connection with rapidly deepening lows, it is known that the change in mass in the stratosphere contributes as much to the local surface pressure change as do the tropospheric changes in density, if not more. Warming is frequently observed in the stratosphere over deepening surface lows, pointing to subsidence in the lower stratosphere. This warming is accompanied by lowering heights of constant pressure surfaces in the lower stratosphere, indicating a decrease in mass at high levels.

Deepening is to a large extent controlled by the mass changes in the upper atmosphere. For example, it has been shown that the lower two-thirds (below about 300-mb level) of the central column become colder and denser as the storm deepened, while the upper one-third of the column became warmer. The upper mass decreases by an amount sufficient to counteract the cooling in the lower layers plus an additional amount to deepen the low. The preferred region for deepening of lows was concluded to be the top third of the atmospheric column or, roughly, the stratosphere. (See fig. 9-9.).

Using the Current 500-Mb Chart

Facsimile transmissions currently contain prognostic 500-mb height contours which can be

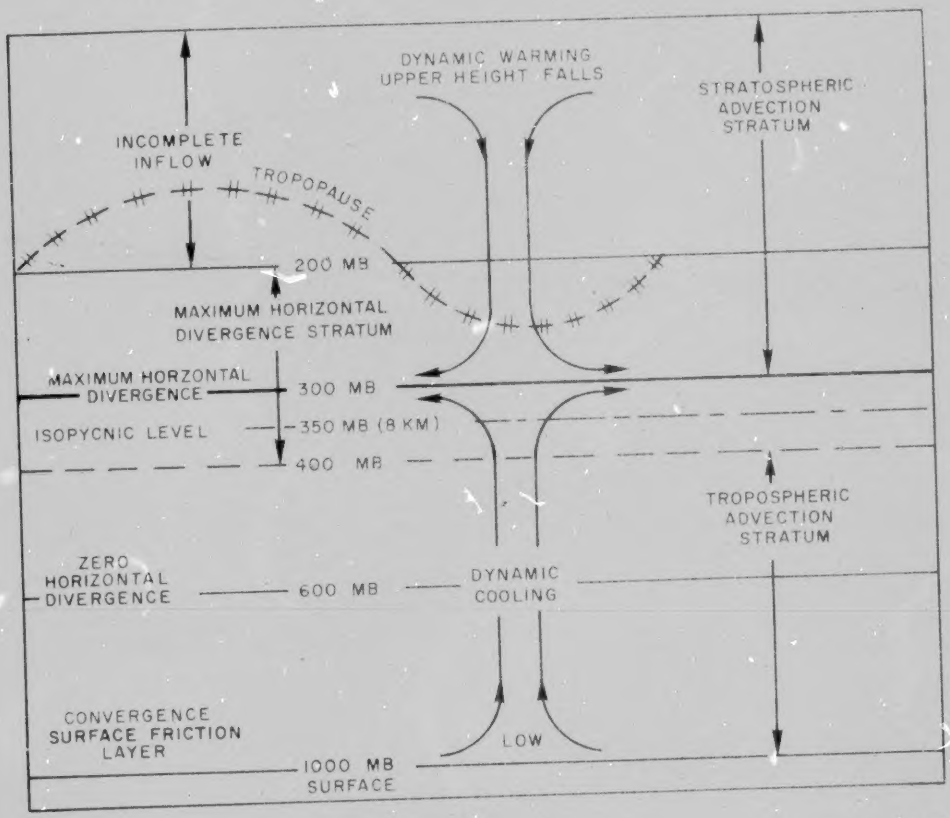
DYNAMIC WARMING
UPPER HEIGHT FALLS
STRATOSPHERIC ADVECTION STRATUM
INCOMPLETE INFLOW
TROPOPAUSE
200 MB
MAXIMUM HORIZONTAL DIVERGENCE STRATUM
MAXIMUM HORZONTAL DIVERGENCE
300 MB
ISOPYCNIC LEVEL
350 MB (8 KM)
400 MB
TROPOSPHERIC ADVECTION STRATUM
ZERO HORIZONTAL DIVERGENCE
600 MB
DYNAMIC COOLING
CONVERGENCE SURFACE FRICTION LAYER
LOW
1000 MB SURFACE

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Figure 9-9.—Vertical circulation over developing low.

used in making predictions of advective changes, thickness patterns, and subsequent changes to the surface pattern.

Deepening Relative to Weather Types

Weather types were discussed previously under the movement of surface lows. This method can also be used to forecast changes in intensity of pressure systems as each system or type has its own average movement plus average deepening or filling.

Deepening of Lows in Relation to Normal Track

Storms whose tracks deviate to the left of the normal track frequently deepen. In general, the normal track of a storm is parallel to the upper flow. If a low deviates to the left of normal, it crosses upper contours (assuming an undisturbed upper current) and becomes superimposed by less mass aloft, resulting in deepening of the low. As long as this crossing of upper contours is unaccompanied by sufficient compensatory cooling at the surface low center, the storm will deepen.

Relation between Deepening Lows and Movement

It is believed that there is little basis for the rule that deepening storms move slowly and filling storms move rapidly. The speed of movement of a low, whatever its intensity, is dependent upon the isallobaric gradient. The magnitude of the surface isallobaric gradients depends upon the low-level advection, the magnitude of the upper-level height changes, and the phase relation between the two.

Using Satellite Photographs

High Resolution Infrared (HRIR) and Advanced Vidicon Camera Systems (AVCS) photographs provide the forecaster with an aid in forecasting the deepening of surface low pressure systems.

In the following series of illustrations (figures 9-10 through 9-15) both AVCS and HRIR photographs show the cloud pictures over a 60 hour period that depict the deepening of a low pressure system.

Figure 9-10, the AVCS picture for noon local time shows a large cloud mass with a low level vortex centered near A. A frontal band, B, extends to the southwest from the large cloud mass. The beginning of a dry tongue is evident. An interesting cloud band, C, which appears just north of the cloud mass is of about the same brightness as the major cloud mass.

The HRIR scan for midnight, figure 9-11, shows the further development of the vortex with penetration of the dry tongue. Low-level circulation is not visible, but the brightness (temperature) distribution differs from the visual picture. The detail within the frontal band is apparent, with a bright cold line, DE, along the upstream edge of the band. This is believed to be the cirrus generated by the convection near the polar jet stream. The cloud band, C, from the previous (daylight) picture is seen in the IR as composed of lower clouds than would be anticipated from the video.

By the second noon (fig. 9-12), the vortex is clearly defined, but again the spiral arm of the frontal band is nearly saturated, with a few shadows to provide detail on cloud layering. The NMC operational surface anlaysis during this period shows that the cyclone has deepened.

Figure 9-13, the picture for the second midnight, shows the coldest temperatures form a hooked shaped pattern with the highest cloudiness still equatorward of the vortex center at F. The flat gray area, G, to the southwest suggests that the area is composed of cells. The granular gray-to-light gray temperature within the dry tongue suggests cells consisting of cumulus formed from stratocumulus, and small white blobs indicating cumulus congestus.

The AVCS picture for the third noon, Figure 9-14, shows the vortex to be tightly spiraled, indicating a mature system. The frontal band is narrower than it was 24 hours earlier, with some cloud shadows present to aid in determining the cloud structure. Surface analysis indicates that the lowest central pressure of the cyclone was reached approximately 6 hours prior to this picture.

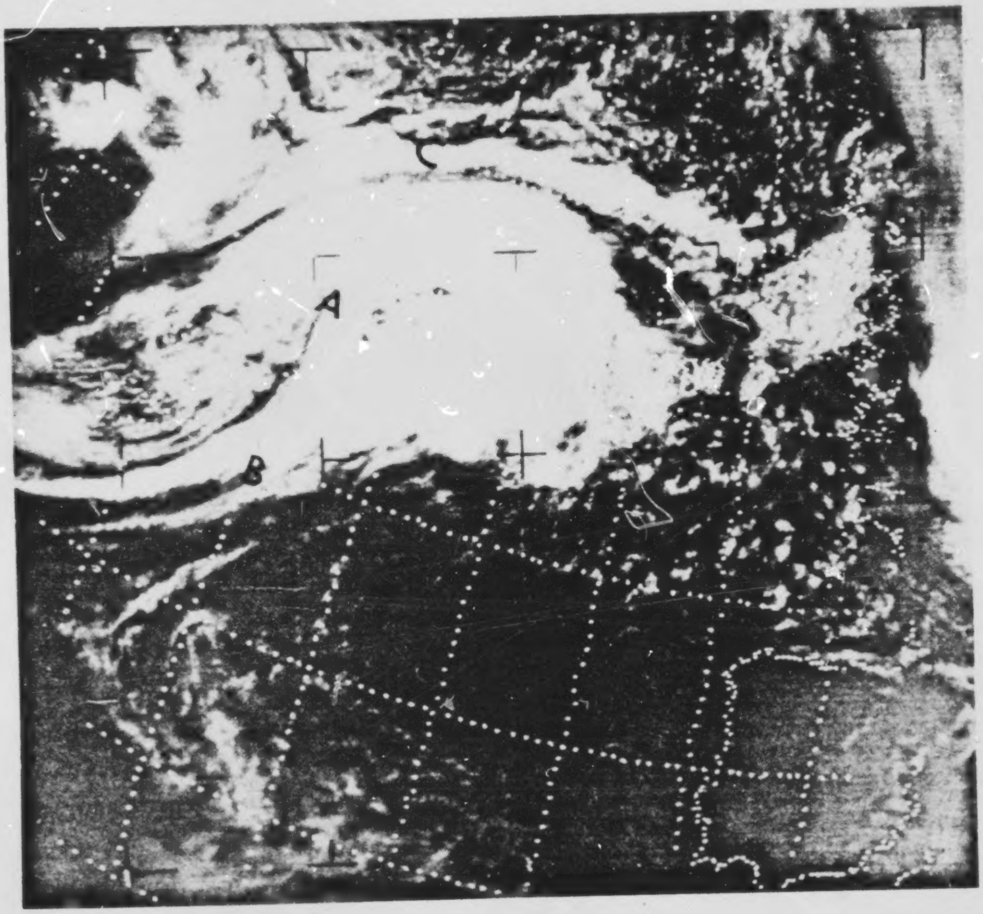

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Figure 9-10.—AVCS, local noon, first day.

The final photograph in this series, figure 9-15, shows the coldest temperatures completely surround the vortex. The frontal band also shows the segmented nature of the active weather areas within the band. A typical vorticity center at H shows the cold temperatures of cumulus congestus and cumulonimbus cloud tops. The same vorticity center is apparent in the previous (daylight) picture, figure 9-14, west of the frontal band.

The forecaster may use the following general conclusions in adapting satellite photographs to forecasting the change in intensity of surface cyclones.

1. The use of nighttime IR data together with daytime data yields 12 hour continuity with respect to cyclone development or decay.

2. A hook-shaped cloud mass composed of cold temperatures, (high cloud tops) indicates an area of strong upward vertical motion with attendant surface pressure falls.

3. As the dry tongue widens, the cyclone continues to deepen.

4. When high or middle clouds completely surround the vortex center, the cyclone has reached maturity and can be expected to fill. This generally indicates the advection of cold, dry air into the cyclone has ceased.

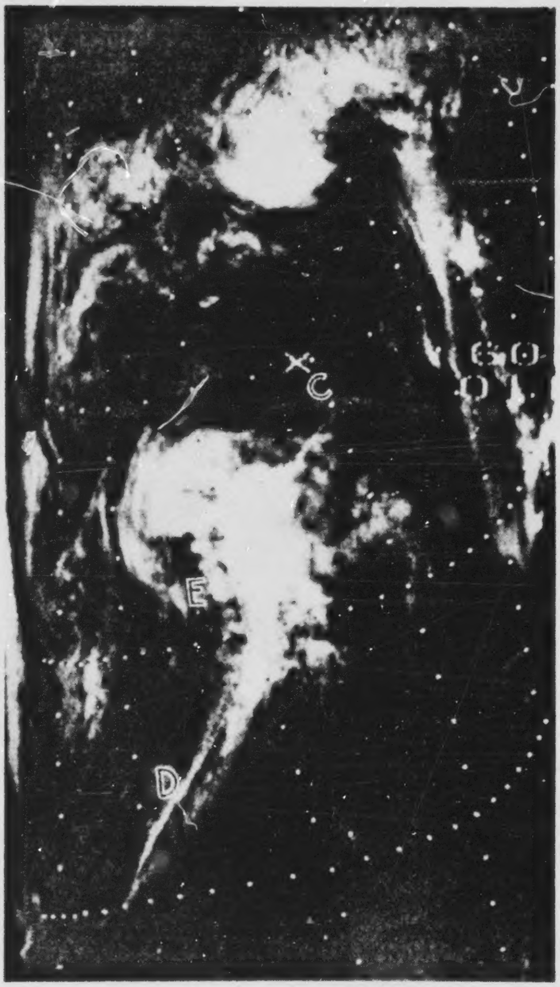

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Figure 9-11.—IR, local midnight, first night.

FORECASTING THE INTENSITY OF HIGHS

Anticyclogenesis Indicators

In the case of developing dynamic anticyclones, it is almost invariably observed that cooling takes place at about 200 mb and above. This cooling is due to ascent of air, resulting from convergence in the 400- to 200-mb stratum. Incomplete outflow at very high levels causes piling up of air above fixed upper levels, resulting in high-level pressure rises. At the same time warming occurs in the lower troposphere. This warming sometimes occurs very rapidly in the lower troposphere above the surface levels, which may remain quite cold. A warming of 10°C per day at the 500-mb level is not unusual. Such a rate of warming is not entirely due to subsidence but probably has a considerable contribution from warm advection. However, continuity considerations suggest that the convergence in the 400- to 200-mb stratum produces some sinking and adiabatic warming in the lower troposphere.

Thus, in the building of anticyclones, there must be a piling up of air at high levels due to horizontal velocity convergence in the 400- to 200-mb stratum, which results in the stratospheric cooling observed with developing anticyclones. Insufficient outflow at very high levels results in an accumulation of mass. This is roughly the mechanism thought to be responsible for the development of high-pressure systems. The high-level increase of mass overcompensates the low tropospheric decrease of density, and the high-level effect thus determines the sign of increase of pressure at the surface when highs are intensifying. (See fig. 9-16.)

The development of anticyclones appears to be just the reverse of the deepening of cyclones. Outside of cold source regions and frequently in cold source regions, high-level anticyclogenesis appears to be associated with an accumulation of mass in the lower stratosphere accompanied by cooling. In many cases this stratospheric cooling may be advective, but more frequently the cooling appears to be clearly dynamic; that is, due to ascent of air resulting from horizontal convergence in the upper troposphere.

Studies of successive soundings accompanying anticyclogenesis outside cold source regions show progressive warming throughout the troposphere. This constitutes a negative contribution to anticyclogenesis. In other words, outside of cold source regions, during anticyclone development, the decrease in DENSITY in the troposphere is overcompensated by an increase in

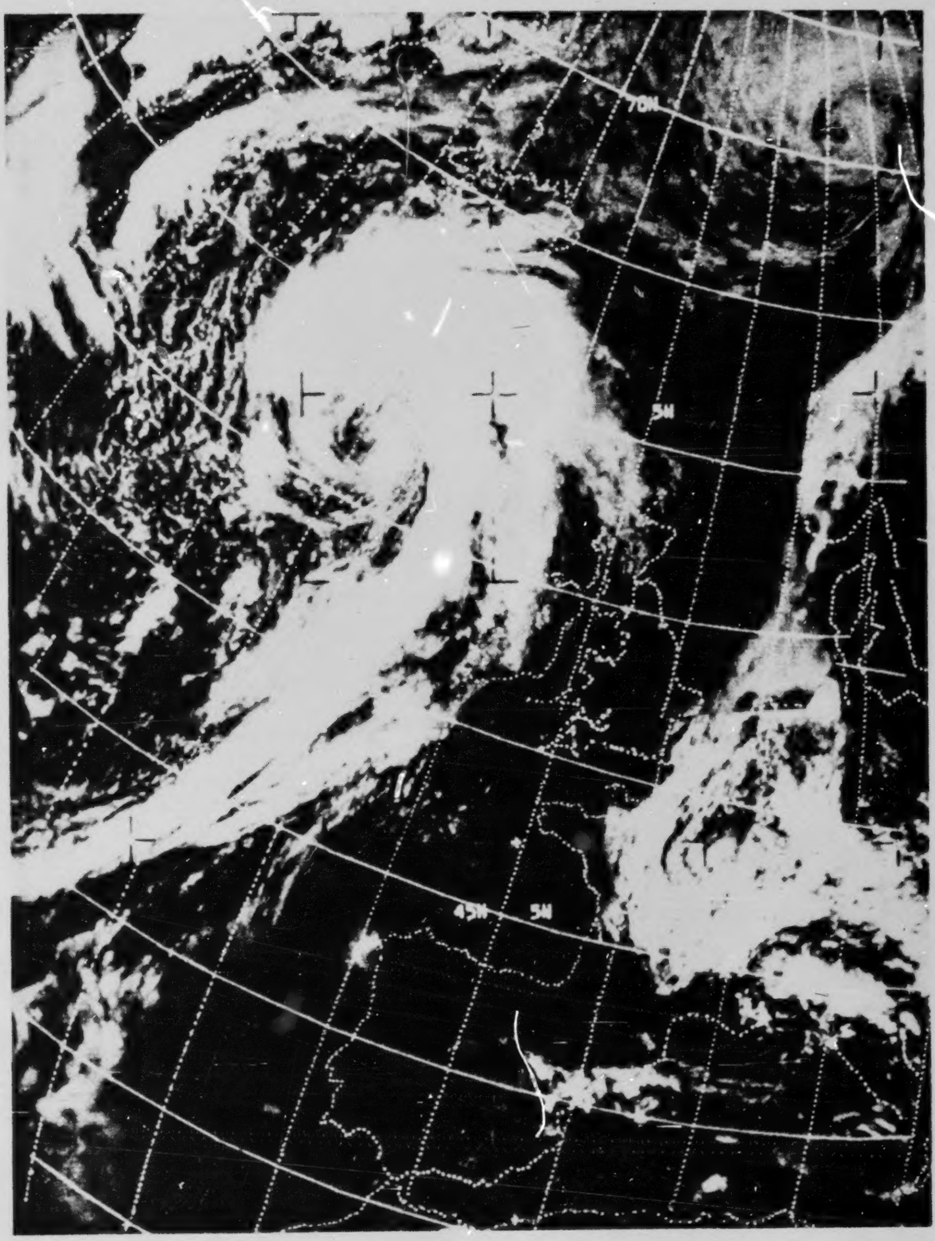
70N
5W
45N
5W

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Figure 9-12.—AVCS, local noon, second day.

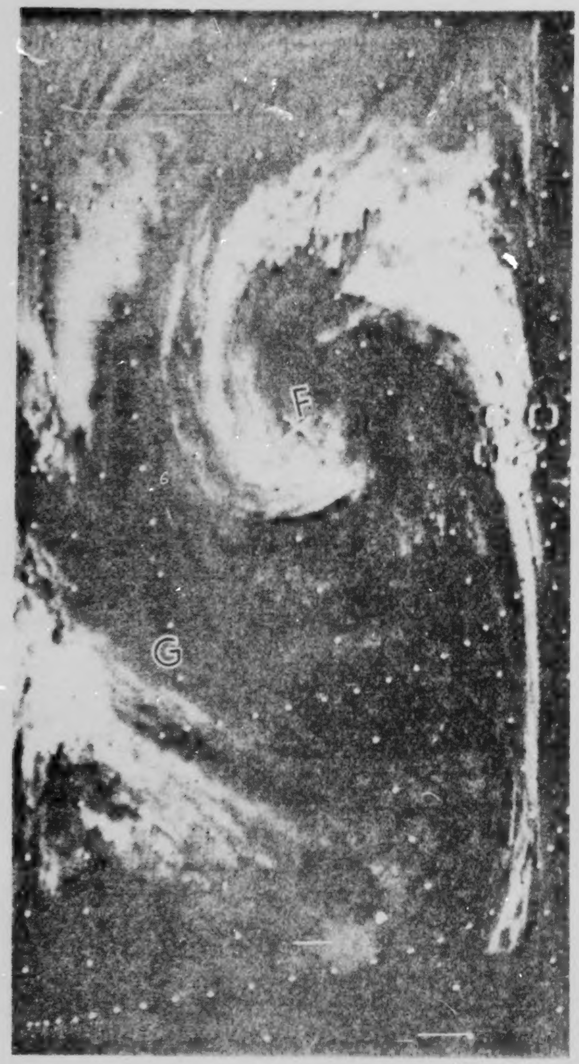

Figure 9-13.—IR, local midnight, second day.

mass, and generally accompanied by cooling in the stratosphere. This is analogous to the deepening of lows where the decrease in mass, generally accompanied by warming at high levels, overcompensates the cooling in the troposphere.

The evidence, therefore, indicates that high-level changes, undoubtedly due to dynamic mechanisms in the upper troposphere, are largely responsible for deepening and filling of surface pressure systems. This fact is of considerable prognostic value if the dynamic processes which induce these mass and density changes can be detected on the working charts.

Progging Intensity of Highs

Intensification of surface highs is indicated and should be progged when cold advection is occurring in the stratum between 1,000 mb and 500 mb when either no height change is occurring (or progged) at 500 mb or when convergence is indicated at and above 500 mb or both, and when the cold advection is increasing rapidly. A high also increases in intensity when the 3-hourly rises are occurring near the center and in the rear quadrants of the high. When a moving surface high which is not subjected to heating from below is associated with a well-defined upper ridge, the change in intensity is largely governed by changes in intensity of the upper-level ridge.

Weakening of surface highs is indicated and should be progged when the cold advection is weakening or is replaced by warm advection in the lower tropospheric stratum with either no height change at 500 mb or when divergence is occurring (or progged) at and above 500 mb or both are occurring at the same time.

When warm low tropospheric advection is coupled with convergence aloft, or when cold low tropospheric advection is coupled with divergence aloft, the contribution of either may be canceled by the other, and a quantitative estimate of the effects of each must be made before a decision is made.

When surface pressure falls occur near the center and in the forward quadrants of the high, the high will decrease in intensity.

When a cold high which is moving southward heats from below, it will decrease in intensity, unless the heating is compensated by intensification of the ridge aloft.

The amount of intensification can be determined by correlating the contributions of height change at the 500-mb level as progged and the change in thickness as advected.

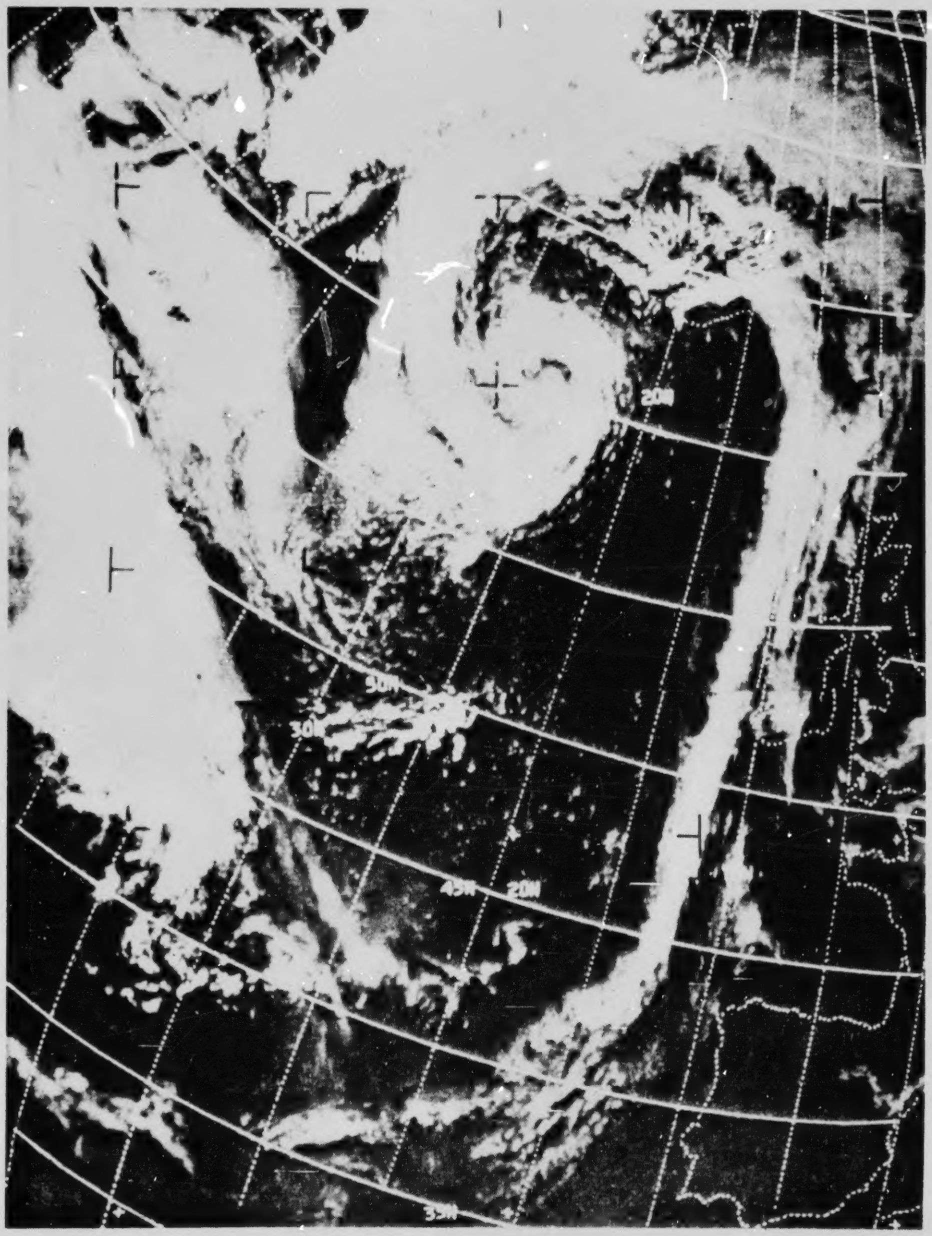
40N
20W
50N
30W
45N
20W
35N

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Figure 9-14.—AVCS, local noon, third day.

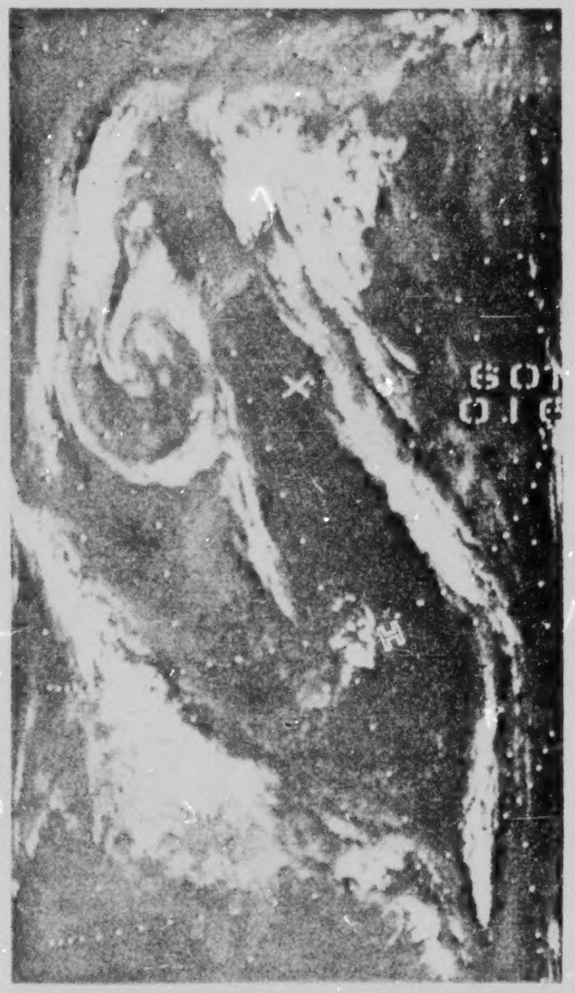

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Figure 9-15.—IR, local midnight, third night.

FORECASTING THE MOVEMENTS AND INTENSITY OF FRONTS

MOVEMENT OF FRONTS

Fronts are ordinarily progged after the pressure systems have been progged. However, there may be cases for short range progging where movement of all of the systems is unnecessary and you are only concerned with a short interval of time. Extrapolation is perhaps the simplest and most widely used method of moving fronts for short periods. When moving fronts for longer periods than a few hours, other considerations must be taken into account.

In this section of the chapter two methods for extrapolation are covered along with the Geostrophic Wind Method and the consideration of upper air influences.

Extrapolation

When fronts are moved by extrapolation, they are merely moved in time and space on the basis of past motion. Of course such factors as occlusion, genesis, dissipation, change in position, intensity of air masses and cyclones, and orographic influences are taken into account. Adjustments to the extrapolation frontal positions are made on the basis of the above considerations.

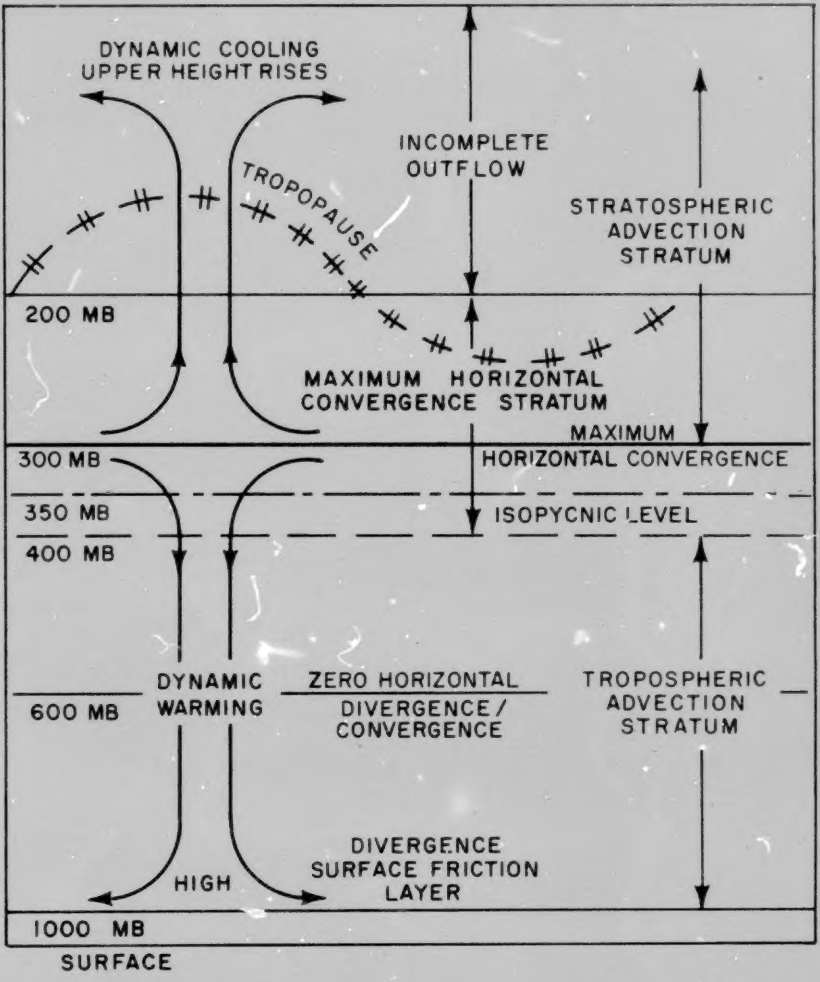
Aerographer's Mates should keep in mind that fronts in themselves do not possess the necessary energy to move, but that the adjacent air masses and associated cyclones drive the fronts.

You should also keep in mind the upper air influences on fronts; for example, the role the 700-mb winds play in the movement and modification of fronts. Finally, it should be remembered that past motion is not a guarantee to future movement, and the emphasis is on considering the changes indicated and incorporation of these changes into a modified extrapolated movement.

CONTROL LINE METHOD OF EXTRAPOLATION.—This method of extrapolation gives a more scientific basis to extrapolation of frontal systems. It could also be applied to cloud systems. The method was adapted from several available methods and published in the Air Weather Service Manual, the Local Area Surface Chart, Its Preparation and Use, AWSM 105-35.

Geostrophic Wind Method

Fronts are progged by the geostrophic wind method at two levels at several points along the front—at the surface and at the 700-mb level. The basic idea is to determine the component of the wind at the surface and aloft which is


DYNAMIC COOLING
UPPER HEIGHT RISES
INCOMPLETE OUTFLOW
TROPOPAUSE
STRATOSPHERIC ADVECTION STRATUM
200 MB
MAXIMUM HORIZONTAL CONVERGENCE STRATUM
MAXIMUM HORIZONTAL CONVERGENCE
300 MB
350 MB
ISOPYCNIC LEVEL
400 MB
DYNAMIC WARMING
ZERO HORIZONTAL DIVERGENCE/ CONVERGENCE
TROPOSPHERIC ADVECTION STRATUM
600 MB
DIVERGENCE SURFACE FRICTION LAYER
HIGH
1000 MB
SURFACE

AG.572

Figure 9-16.—Vertical circulation over developing high.

normal (perpendicular) to the front and therefore drives the front. Determination of the component normal to the front is made by triangulation.

The procedure is as follows:

1. Select between two to four points along the front where a regular and reliable pressure gradient exists, and determine the geostrophic wind by use of the geostrophic wind scales rather than the observed wind. The wind speed is determined a short distance behind a cold front and a short distance ahead of the warm front where a representative gradient can be found. The dots on the isobars in figure 9-17 serve as a guide to the proper selection of the geostrophic wind.

2. Draw a vector toward the front, parallel to the isobars from where the geostrophic wind was determined. The vectors labeled "y" in figure 9-17 illustrate this step.

3. Draw a vector perpendicular to the front originating at the point where the "y" vector

intersects the front and label this vector "x" as illustrated in figure 9-17.

4. At a convenient distance from the intersection, along the "x" vector, construct a perpendicular to the "x" vector, letting it intersect the "y" vector. This is line c in figure 9-17.

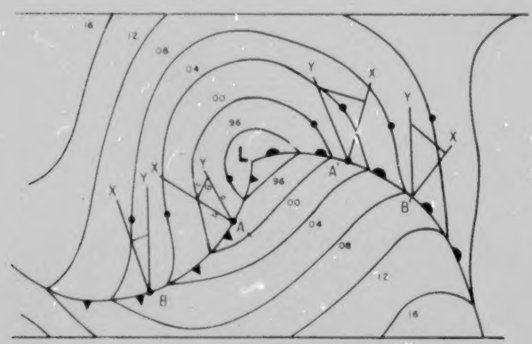


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Figure 9-17.—Geostrophic wind method.

5. The angle formed at the intersection of the "y" vector and the perpendicular originating from the "x" vector is labeled θ (theta). Measure angle θ to the nearest degree with a protractor, and determine the value of its sine by using trigonometric tables or a slide rule.

6. Let side a of the right triangle formed in step 4 represent the value of the geostrophic wind obtained in step 1 and call it "Cgs." Solve the triangle for side b by multiplying the sine of θ by the value of Cgs. The resulting value of b is the component of the wind normal to the front, giving it its forward motion. The formula is

$$b = Cgs \times \sin\theta$$

In a sample problem, if the Cgs was determined to be 25 knots and angle θ to be 40°, b is 19.1 knots, since the sine of angle θ is 0.643.

The Aerographer's Mate can see, however, that the components normal to the front should be equal on both sides of the front and that in reality it would matter very little where the component is computed in advance of or to the rear of the front. In cold fronts the reason why the component to the rear is chosen is that this flow, this air mass, is the one supplying the push for the forward motion and that in the case of warm front the receding cold air mass under the warm front determines the forward motion, because the warm air mass merely replaces the cold air and does not displace it.

In the foregoing discussion we have neglected the effects of cyclonic and anticyclonic curvature in the isobars and the effect of vertical motion along the frontal surfaces. The Aerographer's Mate can readily see that upglide motion along the frontal surfaces reduces the effective component normal to the front and furthermore that cyclonic curvature in the isobars, because it indicates convergence in the horizontal (therefore divergence or mass reduction in the vertical), reduces the effective component normal to the front. For these reasons, the component normal to the front is reduced at the surface only by the following amounts for the different types of fronts and isobaric curvature:

Slow moving cold front, anticyclonic curvature	0%
Fast moving cold front, cyclonic curvature	10-20%
Warm front	20-40%
Warm occluded fronts	20-40%
Cold occluded fronts	10-30%

If the pressure gradient is forecast to increase, decrease the component by the least percentage.

If the pressure gradient is forecast to decrease, decrease the component normal to the front by the highest percentage value.

If the pressure gradient is forecast to remain static, decrease the component normal to the front by the middle percentage value as listed above.

Upper Air Influence on the Movement of Fronts

A number of the rules relating the upper air contours to the movement of fronts were discussed in chapter 5. You saw that a slow moving cold front has parallel contours behind the front, and in a fast moving cold front the contours were at an angle to the front even at times normal to the front.

Some additional rules are stated below:

1. During periods of strong, continued westerly flow aloft (high index) over North America, surface fronts move rapidly eastward. A rule of thumb that can be used with this situation is that the front will move eastward at a speed which is 50 percent of the 500-mb flow and 70 percent of the 700-mb flow.
2. Cold fronts associated with cP outbreaks are closely associated with the depth of the northerly winds aloft. The following relationships are evident: For cP air to push southward into the Great Basin from British Columbia, strong northerlies must exist to at least 500 mb over the area; for cP air to push southward into the Gulf of Mexico to provide extensive northers, it is recognized that northerlies and northwesterlies must exist or be expected at 500 mb as far south as Texas; for cP air to push southward over Florida to Cuba, northerlies must exist at 500 mb as far south as the Gulf States.

FORECASTING THE INTENSITY OF FRONTS

Frontogenesis

Fronts intensity when one of the following three conditions or a combination of them occurs:

1. The mean isotherms (thickness lines) become packed along the front.
2. The fronts approach deep upper troughs.
3. Either or both air masses move over a surface which strengthens their original properties.

The formation of new fronts is a phenomenon rarely observed. Air mass boundaries seem to be perpetually present on the various weather charts. Frontogenesis occurs when two adjacent air masses exhibit different temperature and density, and prevailing winds bring them together. This condition, however, is the normal permanent condition along the polar front zone; therefore, the polar front is semipermanent.

Generation of a new front or strengthening of an existing front occurs during the winter months along the eastern coasts of the American and Asian Continents. When the continent is much cooler than the bordering ocean, frontogenesis may be progged when the air mass moving out onto the ocean has no distinct or a weak boundary.

Frontolysis

Weakening or dissipation of fronts occurs when:

1. The mean isotherms become more perpendicular to the front or more widely spaced.
2. They move out of a deep pressure trough.
3. Either or both air masses modify.
4. They meet with orographic barriers.

PROGNOSTICATING ISOBARS

Isobars may be constructed on the surface prog either by computing the central pressures for the high and low centers and numerous other points on the surface prog chart and drawing the isobars to fit the combination of highs, lows, fronts, and the computed pressures or by moving the isobars in accordance with the surface tendencies and indications.

USING THICKNESS PROGS

The computational procedure for constructing prognostic isobars using the thickness prognosis is as follows:

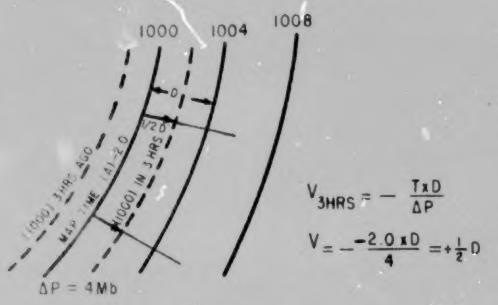
1. At a selected point, determine the difference between the present 500-mb height and the progged 500-mb height. A progged rise at the 500-mb level is positive; a progged fall, negative.
2. At the same selected point, determine the difference between the current and the progged thickness value, a progged increase in thickness being positive in value and a progged decrease in thickness being negative.
3. Subtract algebraically the difference of the progged thickness value from the progged difference of the 500-mb level.
4. Convert this difference in feet to millibars by letting 60 meters equal 7 1/2 mb, and assign the proper sign.

5. Add (or subtract if the value is negative) this value to the current sea level pressure. This is the progged sea level pressure.

6. Repeat steps 1 through 5 for all the points selected, and sketch the isobars.

USING BAROMETRIC TENDENCIES

The variations of the changes in the pressure distribution from one chart depend upon the barometric tendencies. Having the pressure tendency at point A on the 1,000 isobar, its displacement along an axis perpendicular to the isobar can be readily determined. (See fig. 9-18.)


1000
1004
1008
D
1/2D
(1000) 3HRS AGO
MAP TIME (A) -2.0
(1000) IN 3 HRS
ΔP = 4Mb
$V_{3HRS} = -\frac{T \times D}{\Delta P}$
$V = -\frac{-2.0 \times D}{4} = +\frac{1}{2}D$

AG.574

Figure 9-18.—Movement of isobars using the tendency formula.

The tendency at A is -2.0 mb, which means the pressure at A was 1,002 mb 3 hours ago. Since the pressure at A has dropped 2.0 mb during the last 3 hours, the 1,002 isobar will be displaced toward isobars representing higher pressures. Such a movement must take place, for with the same drop in pressure during the next 3 hours, the barometer at A would read 998 mb. Assuming the same drop would occur all along the 1,000 isobar, it would be moved to a point halfway between its present position and the present position of the 1,004 isobar.

This example illustrates the formula shown in figure 9-18 where V is the 3 hourly movement of an isobar, T is the local pressure tendency, D is the distance between neighboring isobars, and Δp is the pressure difference between isobars.

In order to use this equation correctly, the following facts must be considered: First, pressure tendencies are given in tenths of a millibar, while the unit isobars are drawn for every 2 or 4 mb. Second the normal period for measuring pressure tendencies is 3 hours so that velocities derived from this formula will be expressed in terms of this unit and whatever unit is used to measure the distance between isobars. Finally, falling tendencies will give positive velocities or movement toward higher valued isobars, and rising tendencies will give negative velocities or movement toward low valued isobars. The velocities calculated are always directed along an axis perpendicular to the isobars.

The above forecast gives the instantaneous velocity of the isobar under consideration. Consequently, to obtain accurate results, the pressure tendency must remain constant during the forecast period. These ideal conditions rarely exist for extended periods of time with the result that forecasts for longer than 24 hours usually become inaccurate in detail because of acceleration.

Extended Weather Forecasts

Long range weather forecasting dates back to World War II, when wartime needs for extended forecasts become necessary. What began in an experimental stage has now become quite routine.

While most routine forecasting does cover a short period of time, generally 24 to 36 hours, the forecaster will, upon occasion, be called upon to provide outlooks for extended periods—up to 7 days or longer.

Facsimile products will be the primary tool of the forecaster in providing extended forecasts. At the present time the NMC provides a package of extended forecast charts. Among these are a 72, 96, and 120 hour sea level prognostic chart, maximum and minimum temperature anomalies for 3, 4, and 5 days, a 5 day mean temperature anomaly, and total precipitation in classes for the period 3-7 days after the forecast day.

Complete information on these charts may be found in the Weather Bureau Forecasters Handbook NO. 1-Facsimile Products.

Of major importance to the forecaster in preparing an extended outlook without the benefit of facsimile prognostic charts is the determination of the type of flow (zonal or meridional) that can be expected to persist during the outlook period. With a zonal flow a steady progression of pressure systems moving regularly can be anticipated. With the change to a meridional flow the introduction of cold polar or Arctic air into the lower latitudes and mixing with the warmer air will create a number of problems to be considered. Among these will be the development of new pressure systems, deepening or filling of present systems, and movement of the systems.

The forecaster should also become familiar with the particular areas that tend to trigger cyclogenesis and/or frontogenesis and utilize this information when preparing his outlook. Many of the climatological publications that are readily available make note of these areas.

Many local area forecaster's handbooks contain detailed information on how various weather situations approaching the station affect the station weather well in advance. These should be used when available.

CHAPTER 10

CLOUDINESS, PRECIPITATION, AND TEMPERATURE FORECASTING

Three of the most important elements in a forecast are cloudiness, precipitation, and temperature. The occurrence of cloudiness and precipitation may be divided into two main categories—that which occurs with fronts, and that which occurs in air masses not associated with fronts.

There are many factors which influence the daily heating and cooling of the atmosphere which, in turn, affect the temperature. Some of these factors are cloudiness, humidity of the air, nature of the earth's surface, wind, elevation, latitude, and the vertical lapse rate of temperature. Cloudiness is quite obvious in its influence on the penetration of insolation to the earth's surface during the daytime and retardation of the net loss of heat by terrestrial radiation at night.

Other factors which affect all three are stability of the lapse rates, modification of air masses, distance from the moisture source, and the topographical influences. These effects are much the same as they are on climate (as discussed in an earlier chapter), but much more on a local scale and over a shorter period of time.

In this chapter we are mainly concerned with the forecasting of middle and upper cloudiness, precipitation, and local temperature. The forecasting of convective clouds and precipitation, and fog and stratus is covered in chapter 11.

CONDENSATION AND PRECIPITATION PRODUCING PROCESSES

CONDENSATION PRODUCING PROCESSES

To convert significant amounts of water vapor into condensation or the sublimation state, the temperature of the air must be reduced close to the dewpoint. Condensation depends upon two variables—the amount of cooling and the relative humidity of the air. Two conditions must be fulfilled if the condensation of water vapor is to occur in the atmosphere: First, the air must be at or near saturation; and second, hygroscopic nuclei must be present. The first condition can be brought about in either of two ways—one by evaporation of more moisture into the air, the other by cooling the air to its dewpoint temperature. The first process can occur only if the vapor pressure of the air is less than the vapor pressure of the moisture source, and results in fog. Cooling is the principal condensation producer. Nonadiabatic cooling processes (for example, radiation and conduction associated with advection) result principally in fog, light drizzle, dew, or frost. The most effective cooling process in the atmosphere is adiabatic lifting of air. It is the only process capable of producing precipitation in appreciable amounts. It is likewise a principal producer of clouds, fog, and drizzle. The meteorological processes which result in vertical motion of air are discussed in the following sections. It should be emphasized that none of the cooling processes are capable of producing condensation by themselves—moisture in the form of water vapor must be present.

PRECIPITATION PRODUCING PROCESSES

Precipitation occurs when the products of condensation and/or sublimation coalesce to form hydrometeors too heavy to be supported by the upward movement of the air. Although the process of coalescence is not fully understood, it is obvious that a large and continuously replenished supply of water droplets or ice crystals (or both) are necessary if appreciable amounts of precipitation are to occur.

Adiabatic lifting of air is brought about by three general methods: orographic lifting, frontal lifting, and vertical stretching (or horizontal convergence). All of these mechanisms are the indirect results of horizontal motion of air and its variation in space.

Orographic Lifting

Orographic lifting is the most effective and intensive of all cooling processes. Horizontal motion is converted into vertical motion in proportion to the slope of the inclined surface. Comparatively flat terrain can have a slope of as much as 1 mile in 20 miles. The greatest extremes in rainfall records, both in amounts and intensities for varying periods, occur at mountain stations. For this reason the forecaster should be thoroughly familiar with the significant features of the terrain.

Frontal Lifting

Frontal lifting is the term applied to the process represented on a front when the inclined surface represents the boundary between two air masses of different densities. In this case, however, the slope ranges from 1/20 to 1/100 or even less. The steeper the front, the more adverse and intense its effects, other factors being equal. These effects were discussed more fully in chapter 6 of this training manual.

Vertical Stretching

Since it is primarily from properties of the horizontal wind field that vertical stretching is detectable, it is more properly called convergence. This term will be used hereafter.

Convergence does not always result in upward motion of air; only if the convergence occurs through an appreciable layer bounded on its lower side by a solid surface, does convection occur. In order for subsidence to occur, there must be a layer of divergence below. Upper divergence associated with lower level convergence is the typical cyclonic circulation. This combination is a very effective precipitation producer.

The examples of convergence and divergence, explained in the foregoing, are definite and clear cut, associated as they are with the centers of closed flow patterns. More subtle, less easily detected types of convergence and divergence are associated with curved, wave-shaped, or straight flow patterns, where the air is everywhere moving in the same general direction. The qualitative variation of convergence and divergence is indicated in figures 10-1, 10-2, 10-3, and 10-4 by means of the following key:

cc marked convergence.
c convergence.
o little or no convergence or divergence.
d divergence.
dd marked divergence.

Arrows have the following interpretations:

→ wind direction and speed.
➞ movement of center.
↝ path of air current.

The left side of figure 10-1 illustrates longitudinal convergence and divergence; the right side illustrates lateral convergence and divergence. Many more complicated situations can be analyzed by separation into these components.

It can be shown mathematically and verified synoptically that a fairly deep layer of air moving with a marked north-south component has associated with it convergence or divergence, depending on the direction and curvature of the path followed by the stream of air. In figure 10-2 the arrows indicate paths of meridional flow in the Northern Hemisphere. In general, equatorward flow is divergent unless turning cyclonically, and poleward flow is convergent unless turning anticyclonically.

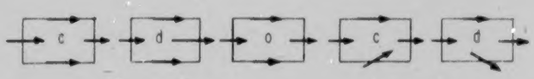


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Figure 10-1.—Longitudinal and lateral convergence and divergence.

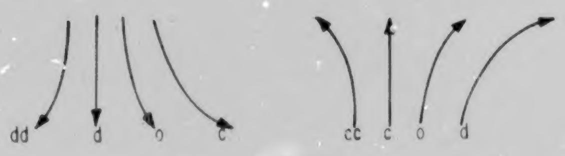

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Figure 10-2.—Convergence and divergence in meridional flow.

The four diagrams of figure 10-3 represent the approximate distribution of convergence and divergence in Northern Hemisphere cyclones and anticyclones. For moving centers, the greatest convergence or divergence occurs on and near the axis along which the system is moving. The diagrams of figure 10-3 show eastward movement, but they apply regardless of the direction of movement of the center.

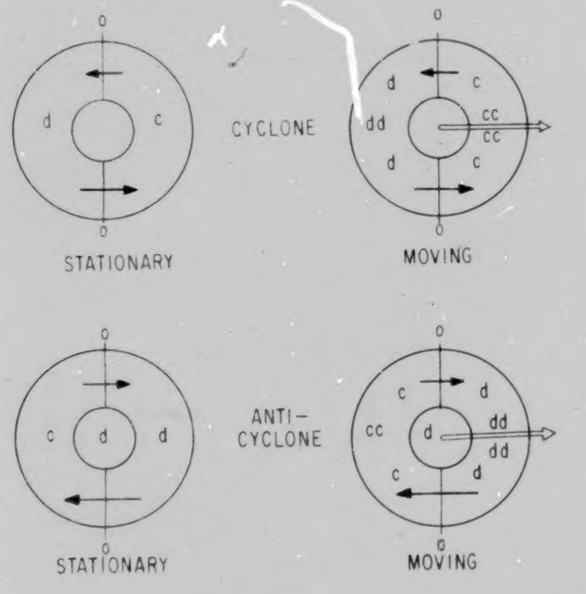

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Figure 10-3.—Convergence and divergence in lows and highs.

Wave-shaped flow patterns are not quite so easily classified with regard to convergence and divergence because the speed at which they travel is often the factor determining the distribution. The most common distribution for waves moving toward the east is illustrated in figure 10-4. There is relatively little divergence at trough and ridge lines, with convergence to the west and divergence to the east of the trough lines. Patterns of this type are more common on upper air charts than on surface charts.

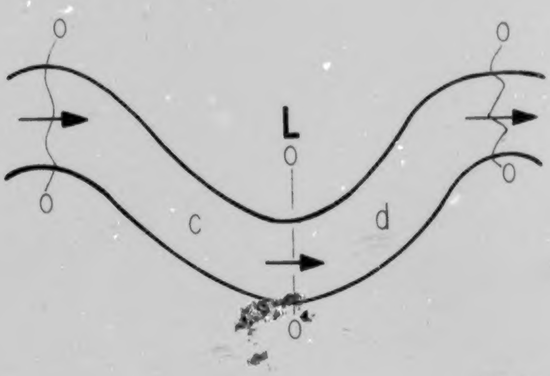

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Figure 10-4.—Convergence and divergence in waves.

Comparatively greater space has been devoted here to convergence because it is the least understood and most difficult to assess of all weather-producing processes. Its areal extent ranges from the extremely local convergence of thunderstorm cells and tornadoes to the large-scale convergence of the broad and deep currents of poleward- and equatorward-moving air masses.

The amount, type, and intensity of the weather phenomena which result from any of the lifting processes described in this section depend on the stability or convective stability of the air being lifted.

It should not be inferred from the discussion of this section that only one of the mechanisms is operative in any particular weather situation. On the contrary, any combination of two, or all three in conjunction, is possible and even probable. For instance, an occluded cyclone of maritime origin moving onto a mountainous west coast of a continent could easily have associated with it warm frontal lifting, cold frontal lifting, orographic lifting, lateral convergence, and convergence in southerly flow. Nearly all fronts have some convergence associated with them.

WEATHER DISSIPATION PROCESSES

Each of the processes described in the preceding section has its counterpart among the condensation-preventing or weather-dissipating processes. Downslope motion on the lee side of orographic barriers results in adiabatic warming, and there is also heating (a nonadiabatic process) of air currents due to insolation from relatively cloudless skies. If the air mass above and ahead of a frontal surface is moving with a relative component away from the front, downslope motion with adiabatic warming will occur. Divergence of air from an area must be compensated for by bringing down air from aloft, which is warmed adiabatically in the process. All these mechanisms have the common effect of increasing the temperature of the air, thus preventing condensation.

These processes likewise occur in combination with one another, but they also occur in combination with the condensation-producing processes. This can lead to situations which require the most careful analysis. For instance, a current of air moving equatorward on a straight or anticyclonically curved path encounters an orographic barrier oriented across its path; if the slope is sufficiently steep or the air is sufficiently moist, precipitation will occur in spite of the divergence and subsidence associated with the flow pattern. The dry, sometimes even cloudless, cold front which moves rapidly from west to east in winter is an example of upper level downslope motion which prevents the air being lifted by the front from reaching the condensation level.

The precipitation process itself opposes the mechanism which produces it, both by contributing the latent heat of vaporization and by exhausting the supply of water vapor if the moisture source is cut off.

FORECASTING FRONTAL CLOUDS AND WEATHER

The cloud and weather systems most difficult to forecast are those associated with new, developmental weather activity, usually that of cyclogenesis. Although there are no known general methods for forecasting cyclogenesis with any great success, it is well known that falling pressure, precipitation, and expanding shields of middle clouds indicate that the cyclogenetic process is operating and, by following these indications, successful forecasts can often be made for 12 to 24 hours in advance. For longer period forecasts, the difficulties increase. Most of the winter precipitation of the lowlands in the midlatitudes is chiefly cyclonic or frontal in origin though convection is involved when the displaced air mass is conditionally or convectively unstable. Cyclones are important generators of precipitation in the Tropics as well as in midlatitudes.

Since the correct cloud and precipitation forecast is a combination of a number of factors, consideration of these factors should be a matter of choice of the individual forecaster. Some of the factors involved and to be considered are listed below, not necessarily in the order of importance:

1. The location of the cyclone and/or front with respect to latitude, geography, terrain, maritime influences, and the surface over which the front or low will move with the modifying influences.
2. The type of low, the type and slope of the front, the stability of the air masses, moisture content and discontinuity, temperature contrast, wind and contour pattern aloft, speed of the low or front, and future movement.

Knowing normal weather patterns with fronts and lows for particular areas will also be helpful. As pointed out before, a thorough understanding of the physical processes by which precipitation develops and spreads is essential to a good precipitation forecast. Another vexing problem to the forecaster is often the separation of frozen from unfrozen types of precipitation. This problem is discussed later in the chapter.

FRONTAL CLOUDINESS AND PRECIPITATION

Cold Front

The study of constant pressure charts in conjunction with the surface synoptic situation is helpful in forecasting cloudiness and precipitation associated with fronts. When the contours at 700 mb are perpendicular to the surface cold front, the band of weather associated with the front is narrow. This is the situation that occurs with a fast moving front. If the front is a slow moving front, the weather and precipitation extends as far behind the front as the winds at the 700-mb level are parallel to the front. In both of these cases the flow at 700 mb also indicates the slope of the front—the region in which the frontal lifting is concentrated. Since the front at 700 mb lies close to the trough line at 700 mb, it is apparent that when the wind at 700 mb is prependicular to the surface front that the 700-mb trough is very nearly above the surface trough; hence the slope of the front is very steep. When the 700-mb flow is parallel to the surface front, the trough lies to the rear of the surface front and beyond the region in which the flow continues parallel to the front. Consequently, the frontal slope is more gradual, and lifting is continuing between the surface and 700 mb at some distance behind the surface frontal position.

Another factor which contributes to the distribution of cloudiness and precipitation in these two situations is the curvature of the flow aloft. Cyclonic flow is associated with horizontal convergence, and anticyclonic flow is associated with horizontal divergence.

Very little weather is associated with a cold front if the mean isotherms are perpendicular to the front. When the mean isotherms are parallel to the front, weather will occur with the front. This principle is associated with the contrast of the two air masses, hence, with the effectiveness of lifting.

Satellite photographs, both Advanced Vidicom Camera Sub System (AVCS) and High Resolution Infrared (HRIR), provide a representative picture of the cloud structure of frontal systems. Active cold fronts appear as continuous well developed cloud bands composed of low, middle, and high clouds. This is caused by the upper wind flow which is parallel, or nearly parallel to the frontal zone (fig. 10-5).

The perpendicular component of the upper winds associated with the inactive cold front cause the cloud bands to appear as narrow, fragmented, or discontinuous. The band of clouds is comprised mainly of low level cumulus and stratiformed clouds, but some cirriform may be present. Inactive cold fronts over water occasionally will have the same appearance as active fronts over land, while over land they may have few or no clouds present. Figure 10-6 depicts the fragmented clouds with an inactive cold front in the lower portion, while a more active cold front cloud presentation is shown in the upper portion.

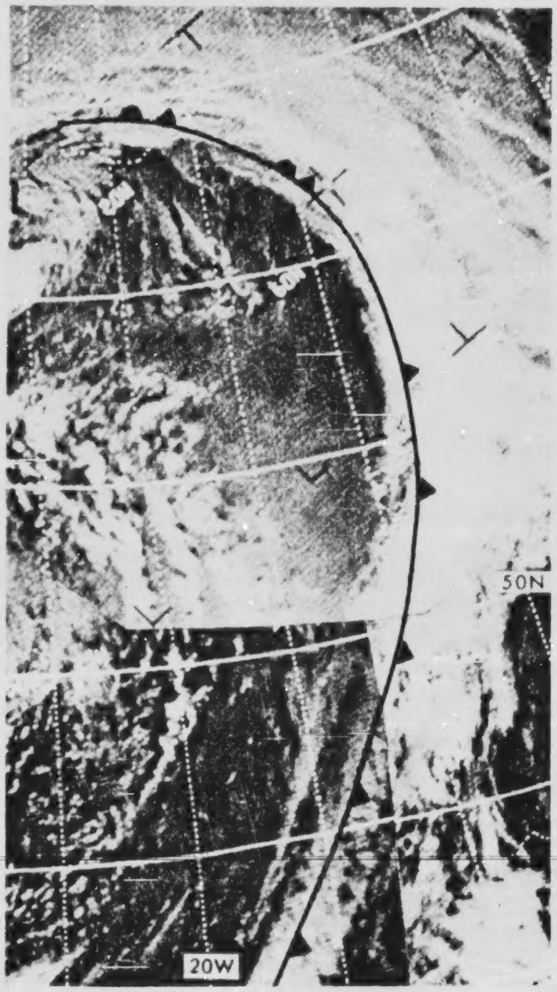

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Figure 10-5.—An active cold front.

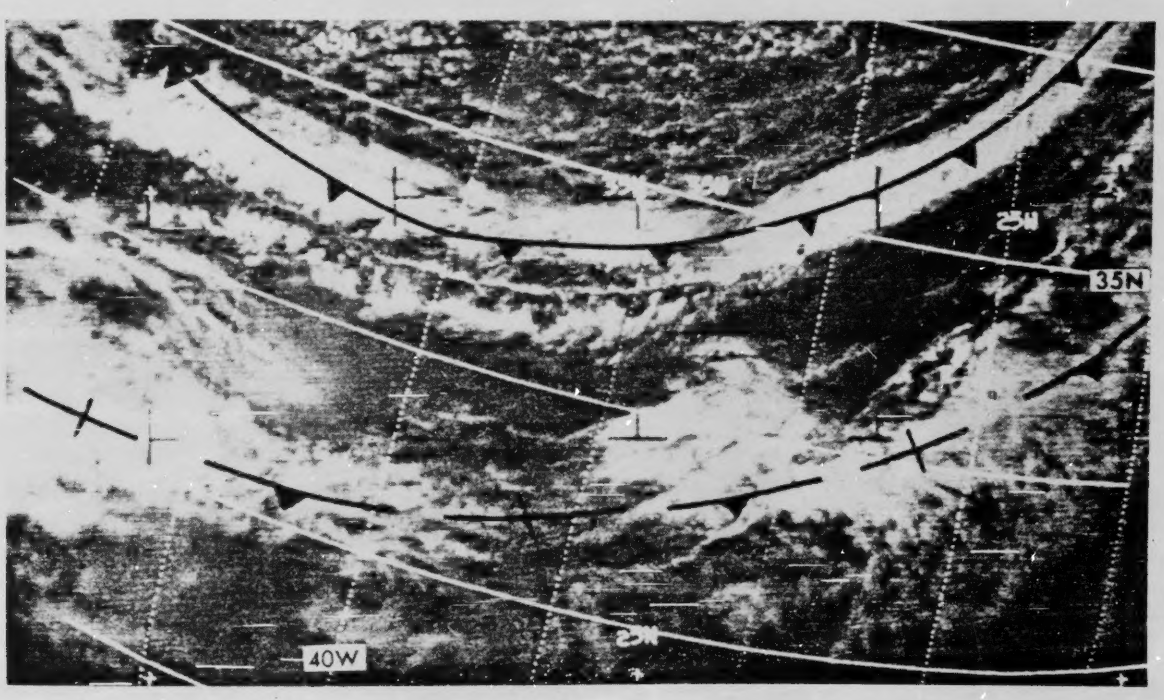

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Figure 10-6.—Inactive and active cold front satellite photograph.

Warm Front

Cloudiness and precipitation occur where the 700-mb flow across the warm front is from the warm to the cold side and is turning cyclonically or moving in a straight line. This implies convergence associated with the cyclonic curvature. Warm fronts are accompanied by no weather and few clouds if the 700-mb flow above them is anticyclonic. This is due to horizontal divergence associated with anticyclonic curvature.

The 700-mb ridge line ahead of a warm front may be considered the forward limit of prewarm front cloudiness. The sharper the ridge line, the more accurate the rule. This, too, is associated with divergence and convergence. The 500-mb ridge line may be considered the forward limit of the cirrus cloud shield.

When the slope of the warm front is nearly horizontal near the surface position and is steep several hundred miles north of the surface position, the area of precipitation is situated in the region where the slope is steep. There may be no precipitation just ahead of the surface frontal position.

A general consensus appears to be that frontal clouds are usually layered; that the high clouds are usually detached from the middle clouds: and that middle clouds are often well layered and many times do not extend very high. It has been found by flights through warm fronts, and other evidence, that the thick solid mass of clouds which extends up to 7 to 8 km in a belt several hundred miles wide along and ahead of the warm front, as depicted by the classical models, is not typical of most fronts.

Active warm fronts are at best, difficult to locate on satellite cloud pictures, while inactive warm fronts cannot be located. An active warm front may be associated with a well organized cloud band, but the frontal zone istelf will be difficult to locate. An active warm front may be placed somewhere under the bulge of clouds that are associated with the peak of the warm sector of a frontal system. The clouds are a combination of stratiform and cumuliform beneath a cirriform covering. (See fig. 10-7.)

It must be remembered that no one condition represents what could be called typical conditions, as each front presents a different situation with respect to the air masses involved. Therefore, each front must be treated as a separate case, using present indications, geographical location, stability of the air masses, moisture content, and intensity of the front to determine its precipitation characteristics.

Orographic Barriers

In general, an orographic barrier increases the extent and duration of cloudiness and precipitation on the windward side of the orographic barrier and decreases it on the lee side of the orographic barrier.

AIR MASS CLOUDINESS AND PRECIPITATION

Surface heating or cooling and movement of the air mass against orographic barriers and movement of the air mass in a cyclonic or anticyclonic path are the important factors in the production of air mass weather.

If an air mass moves so that it is lifted over an orographic barrier and the lifting is sufficient for the air to reach its lifting condensation level, cloudiness of the convective type occurs. If the

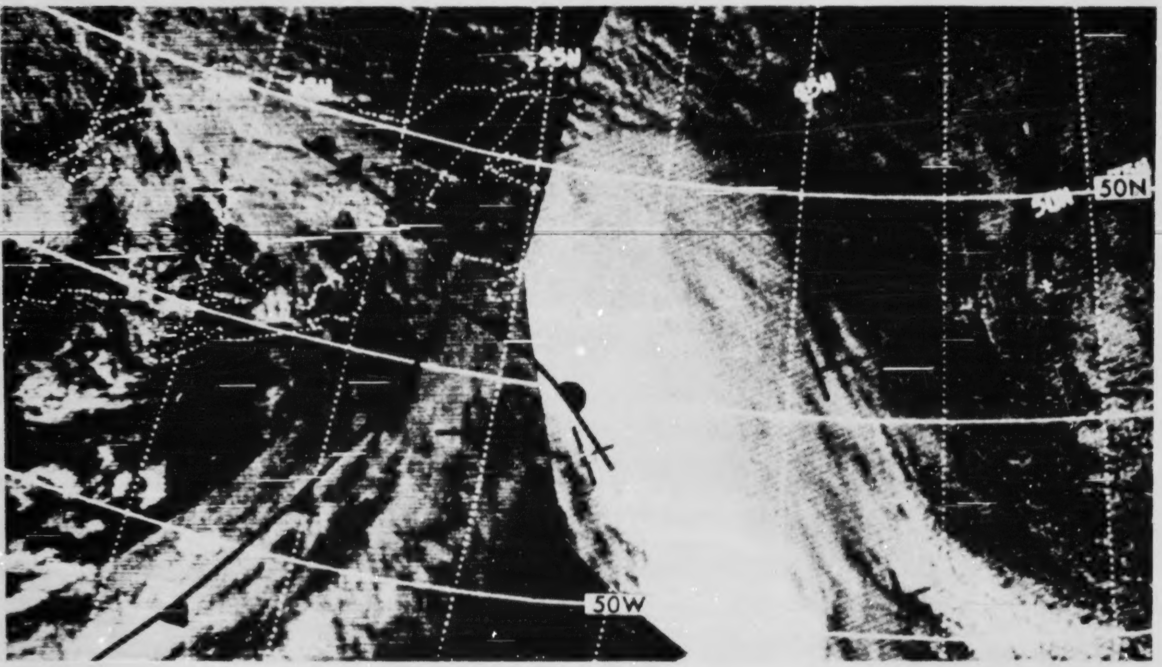

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Figure 10-7.—Warm front satellite cloud photograph.

air is convectively unstable and has sufficient moisture, showers or thunderstorms occur. This is, of course, on the windward side of the barrier.

Curvature of the flow aloft also affects the occurrence of cloudiness and precipitation. Actually the shear should also be considered, but it is seldom considered, since it is difficult to evaluate at levels below the 300-mb level. In a cold air mass, showers and cumulus and stratocumulus clouds are found only in those portions of the air mass which are moving in a cyclonically curved path. In a warm air mass moving with a component from the south, cloudiness and precipitation will be very abundant under a current turning cyclonically or even moving in a straight line. Clear skies occur wherever a current of air is moving from the north in a straight line or in an anticycionically curved path. Clear skies also are observed in a current of air moving from the south if it is turning sharply anticyclonically. Elongated V-shaped troughs aloft have cloudiness and precipitation in the southerly current in advance of the troughs, with clearing at the trough line and behind it. These rules also apply in situations where this type of low is associated with frontal situations.

Cellular cloud patterns (open or closed) as shown by satellite cloud photographs will provide the forecaster with information to identify regions of cold air advection, areas of cyclonic, anticyclonic, and divergent wind flow within air masses, especially behind polar fronts and over oceanic areas.

Open cellular patterns are most commonly found to the rear of cold fronts in cold unstable air. These patterns are made up of many individual cumuliformed cells. The cells are composed of cloudless, or less cloudy, centers surrounded by cloud walls with a predominant ring or U-shape. In the polar air mass the open cellular patterns that form in the deep cold air are predominately cumulus congestus and cumulonimbus. The open cells that form in the subtropical high are mainly stratocumulus, cumulus, or cumulus congestus clusters. In order for open cells to form in a polar high, there must be moderate to intense heating (a large air-sea temperature difference) of the air mass from below. When the cold continental air moves over the water, the moist layer is shallow and capped by a subsidence inversion near the coast. Further downstream the vertical height of the moist layer increases, as does the height of the clouds, and the capping is caused by dry air entrainment. In figure 10-8 the open cells behind a polar front over the North Atlantic indicate cold air advection and cyclonic curvature of the low level wind flow. Vertical thickness of the cumulus at A is small but increases eastward toward B.

Figure 10-9 shows a large area of the subtropical high west of Peru covered with open cells. These are not associated with low level cyclonic flow or steady cold air advection.

Closed cellular patterns are characterized by approximately polygonal cloud covered areas bounded by clear or less cloudy walls. Atmospheric conditions necessary for the formation of closed cells are weak convective mixing in the lower levels and a cap to this mixing. The convective mixing is the result of surface heating of the air or radiational cooling of the cloud tops. This convection is not as intense as that associated with open cells. The cap to the instability associated with these closed cellular cloud patterns is in the form of a subsidence inversion in both polar and subtropical situations. Closed cellular patterns are made up of stratocumulus elements in both the polar and subtropical air masses and may also be accompanied by trade wind cumulus in the subtropical high. Closed cells, when associated with the subtropical highs, are located in the eastern sections of the high pressure area. Closed cells are associated with limited low level instability below the subsidence inversion and extensive vertical convective activity is not likely. Figure 10-10 shows closed cells in the southeastern portion of a polar high near A.

Figure 10-11 shows closed cells in the eastern portion of a subtropical high in the South Pacific. West of A the closed cells are composed of stratocumulus with some clear walls and east of the walls are composed of thinner clouds.

VERTICAL MOTION AND WEATHER

Upward motion is associated with increasing cloudiness and precipitation, subsidence with improving weather. There have been a number

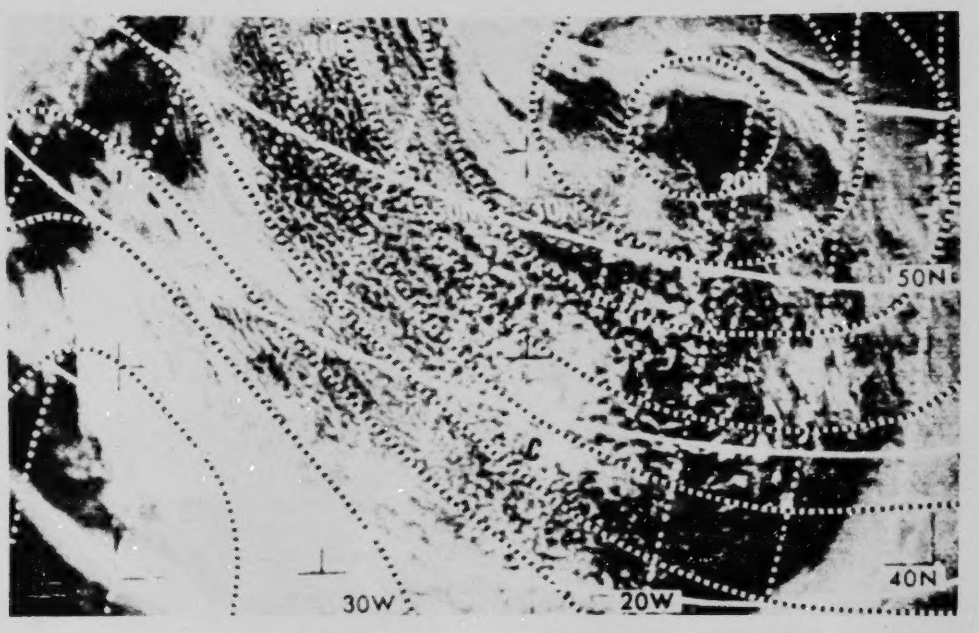

Figure 10-8.—Open cells on a satellite photograph. AG.582

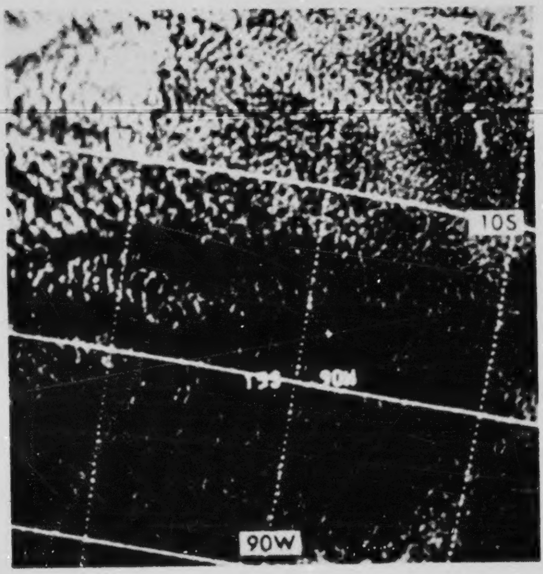

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Figure 10-9.—Open cells in the subtropical high.

of studies of the relationship between precipitation and vertical velocities computed by various techniques. In all cases, the probability of precipitation is considerably higher in the 6 hours following an updraft than following subsidence. Clear skies are most likely following downdrafts.

Vertical motion analysis charts and prognostic charts are currently being transmitted on the National Facsimile Network. The values are computed by Numerical Weather Prediction Methods and are analyzed on the charts. Plus values represent upward motion, and minus values downward motion (subsidence).

A more detailed discussion of the relationship of vertical motion to middle cloudiness precipitation, and convective activity may be found in Vertical Motion and Weather Forecasting, NWRF 30-0359-024.

VORTICITY AND PRECIPITATION

Vorticity has been discussed in detail in chapter 4 of this training manual. We have seen that relative vorticity is due to the effects of

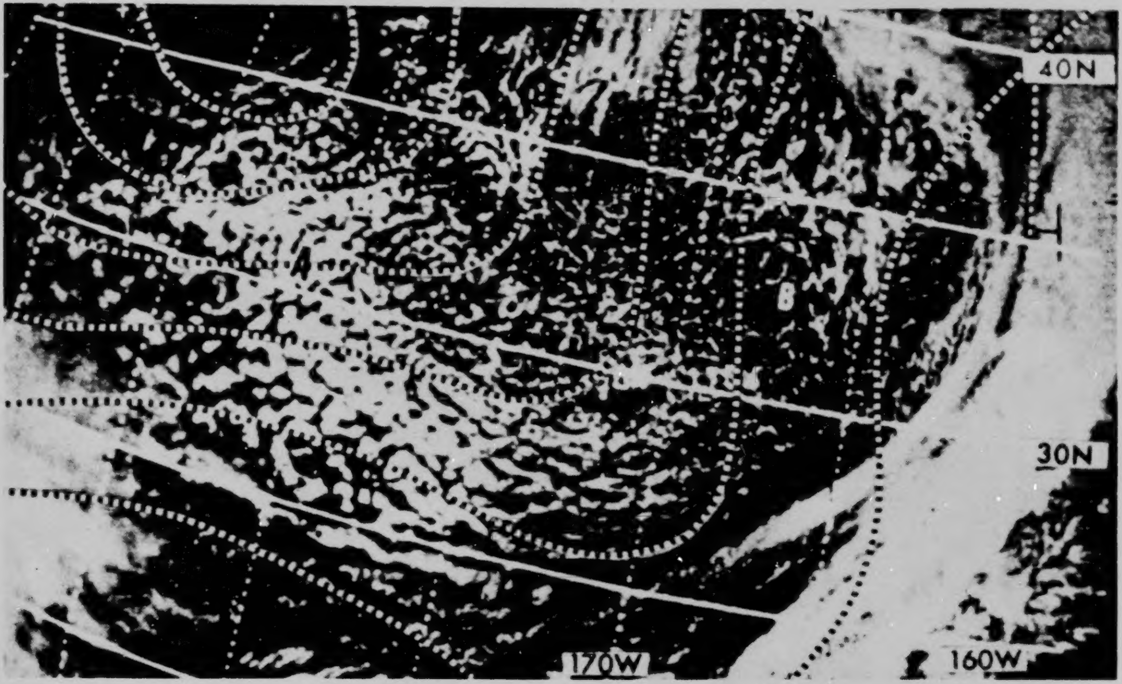

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Figure 10-10.—Closed cells in polar high.

both curvature and shear. Studies have led to the following rule: Cloudiness and precipitation should prevail in regions where the relative vorticity decreases downstream along the streamlines. Fair weather should prevail where it increases along the streamlines.

The fact that both wind shear and curvature must be considered when relative vorticity changes are investigated results in a large number of possible combinations on upper air charts. When both terms are in agreement, we can confidently predict precipitation or fair weather, as the case may be. When the two terms are in conflict, computation must be made before the forecast can be made. Since the former case should include the most potent rain producing situations, actual computations are, in general, unnecessary.

For a full discussion of this problem, it is suggested that the AG refer to Precipitation and Vorticity, AROWA 13-0953-092.

MIDDLE CLOUDS IN RELATION TO THE JETSTREAM

Jets show as much individuality with respect to associated weather as do fronts. Because of the individuality of jetstreams, and also because of the individuality of each situation with respect to humidity distribution and lower level circulation patterns, statistically stated relationships become somewhat vague and are of little value in forecasting. However, the results of one study of visual cloud observations from meteorological research probing of jetstreams found that the layer types of middle clouds were reported mostly to the right (or high-pressure side) of the jetstream axis.

SHORT RANGE EXTRAPOLATION FOR TERMINAL FORECASTING

The purpose of this section of the chapter is to outline several methods which are particularly

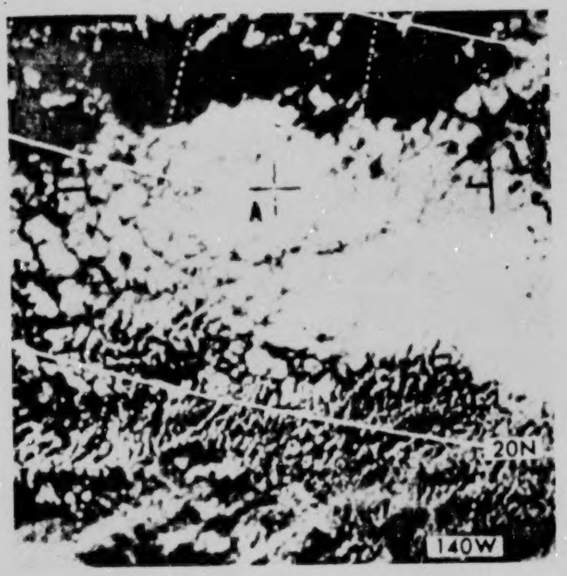

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Figure 10-11.—Closed cells in a subtropical high.

suited for preparing forecasts for periods of 1 to 3 hours. All the techniques presented are based on a single forecasting principle—extrapolation. Extrapolation is the most powerful short range forecasting tool currently available to meteorologists.

Extrapolation is the estimating of the future value of some variable from observations of its present and past values. In its simplest form, as discussed in this section, extrapolation is taken linearly, with only very crude, if any, consideration given to accelerations. The quality and frequency of observations do not generally permit or justify use of more sophisticated techniques for treating accelerations, nor do short range forecasts generally require them. Even the simplest methods are sensitive to the quality and time spacing of the observations as well as the quality of any analysis of these observations.

The hourly sequences obviously cannot provide a complete nor even a wholly adequate picture of the weather under all conditions. When special conditions prevail, the REMARKS section provides opportunity for simplifying the observations. This may be of great importance in bad flying weather. The forecasters and observers should utilize the remarks section as much as possible to this end.

NEPHANALYSIS

Nephanalysis may be defined as any form of analysis of the field of cloud cover and/or type. The potentialities of specialized nephanalysis may seem to be obvious, but in practice, difficulties are encountered. The cloud observations received in synoptic codes permit only a highly generalized and incomplete discription of the actual structure of the cloud systems; the observation stations are usually too far apart to permit a representative picture of the distribution of many features which have a high degree of spatial and time variability; so many parameters are observed that they cannot all be readily analyzed on one chart.

In view of the above limitations, forecasters should approach this problem with an open mind for further experimentation. You should select for analysis only the particular parts of the cloud observations which are important to the intended application, then adjust the chart scale, the degree of detail, and the mode of representation to the character of the data and to the purpose.

At present, few forecasters make full use of the cloud reports plotted on their charts, and often many cloud reports are omitted. Obviously, the first consideration in nephanalysis is to survey what cloud information is transmitted and to make sure that everything pertinent is plotted on the regular charts. For very short range forecast, the charts at 6-, 12-, and 24-hour intervals are apt to be insufficiently frequent for use of the extrapolation techniques explained in this chapter. Either neph charts or surface charts should then be plotted at the intermediate times from 3-hourly synoptic reports or even from hourly sequences. An integrated system of forecasting ceiling, visibility, cloud cover, and precipitation should be considered simultaneously as these elements are physically dependent upon the same synoptic processes.

With the advent of satellite cloud pictures it is rare that hephanalysis form manually plotted data will be performed. Instead, the surface analysis and satellite cloud pictures will be utilized together.

FRONTAL PRECIPITATION

For short range terminal forecasting, the question is often not whether there will be any precipitation, but when will it begin or end; for example, in cases where it has already begun upstream or at the terminal.

This problem is well suited to extrapolation methods. The prediction of new precipitation areas, such as those associated with new wave development, upper troughs, etc., of course, requires other synoptic methods. However, for very short range forecasting, the use of hourly nephanalysis often serves to "pick up" new precipitation areas forming upstream in sufficient time to alert a downstream area. Also, the thickening and lowering of middle cloud (altostratus) decks generally indicate where an outbreak of precipitation may soon occur.

Forecasting the Movement of Precipitation Areas by Isochrones

The areas of continuous, intermittent, and showery precipitation can be outlined on a special large-scale 3-hourly or hourly synoptic chart in a manner similar to the customary shading of precipitation areas on ordinary synoptic surface weather maps. Different types of lines, shading, or symbols can distinguish the various types of precipitation. Isochrones of several hourly past positions of the lines of particular interest can then be added to the chart and extrapolations for several hours made from them if reasonably regular past motions are in evidence. A separate isochrone chart (or acetate overlay) may be easier to use. Lines for the beginning of continuous precipitation are illustrated in figure 10-12. The isochrones for showery or intermittent precipitation usually give more uncertain and irregular patterns which result in less satisfactory forecasts. When large-scale section surface weather maps are regularly drawn, it may be sufficient and more convenient to make all precipitation area analyses and isochrones on these maps.

Precipitation Movement Using a Distance Versus Time Diagram

The idea of plotting observations taken at different times on a diagram which has horizontal or vertical distance in the atmosphere as one coordinate and time as the other is very old and has been used in various forms for diverse purposes by meteorologists for years. The time cross sections (as explained in chapter 12 of this training manual) are a special case of this device, where successive information at only one station is plotted.

On this diagram the horizontal coordinate is an axis line passing from the terminal in question through a series of stations in the direction from which the weather usually approaches, or from which it is expected during the particular situation. This line forms the distance, or x-axis, of the diagram. The positions of the stations are marked off on it at their appropriate scale distances. The time scale (increasing upward) is erected at right angles to the origin of the x-axis. Both coordinate axes have linear scales. Observations for any time and points reasonably near the x-axis may be plotted on this diagram, using the standard plotting models for airways or synoptic reports. The resulting picture has somewhat the character of a plotted synoptic chart and can be analyzed accordingly.

When complete synoptic models are plotted on the x-t diagram, various factors affecting local movement and development of systems can be evaluated subjectively in making the extrapolation. For example, the interrelations of fronts, visibilities, winds, clouds, temperatures, pressures, and precipitation can be readily considered. Moreover, locally important details of the situation missed on the large-scale regular synoptic weather maps can often be spotted and followed on this diagram.

LOWERING OF CEILING IN CONTINUOUS RAIN AREAS

Frontal Situations

The lowering of ceilings with continuous rain or snow in warm frontal and upper trough situations is a familiar problem to the forecaster

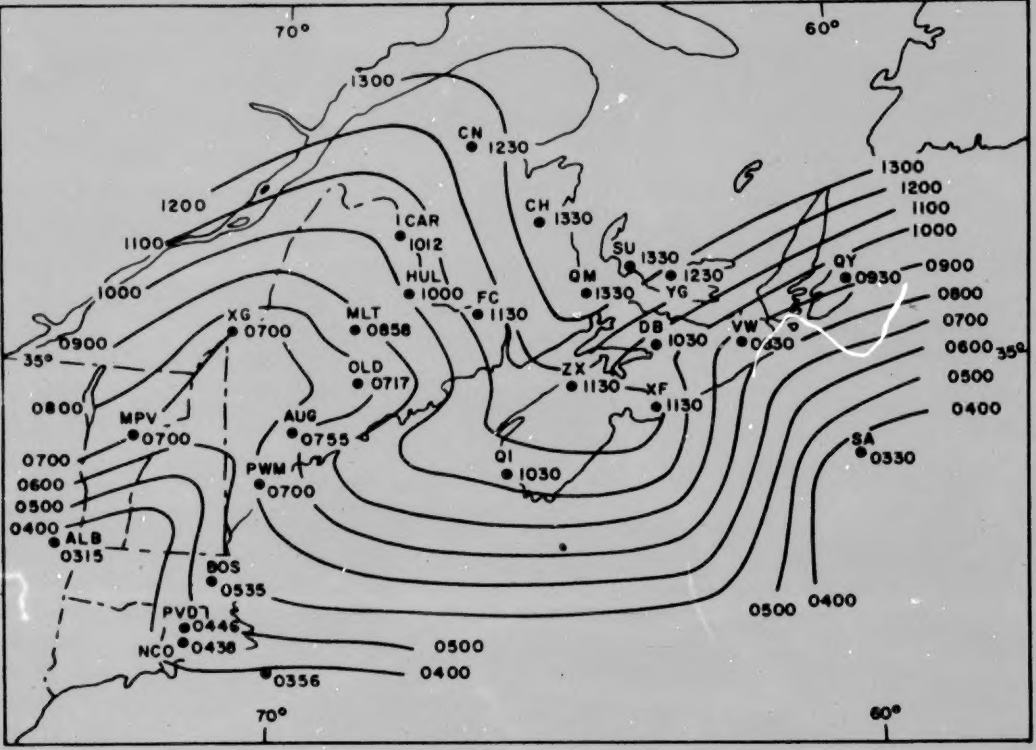
70°
60°
1300
1200
1100
1000
0900
0800
0700
0600
0500
0400
35°
CN 1230
CH 1330
CAR 1012
HUL 1000
FC 1130
SU 1330
QM 1330
1230
YG
QY 0930
XG 0700
MLT 0858
OLD 0717
DB 1030
VW 0930
ZX 1130
XF 1130
MPV 0700
AUG 0755
PWM 0700
QI 1030
SA 0330
ALB 0315
BOS 0535
PVD 0446
NCO 0438
0356
0500
0400
70°
60°

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Figure 10-12.—Isochrones of beginning of precipitation, an early winter situation.

in many regions. In very short range forecasting, the question as to whether or not it will rain and when the rain will begin is not so often the critical one. Rather, the problem is more likely to be (assuming the rain has started) how much will the ceiling lower in 1, 2, and 3 hours, or will the ceiling go below a certain minimum in 3 hours. The visibility in these situations generally does not reach an operational minimum as soon as the ceiling, and fog usually becomes no worse than light. It has been shown that, without sufficient convergence, advection or turbulence, evaporation of rain into calm lower air does not lead to saturation and causes no more than haze or light fog.

At some locations with advection, upslope trajectories may bring in more moist air, where the stratus or dense fog that forms is more or less independent of the rain-evaporation effect. Once stratus has formed, however, extrapolation of its displacement may be quite successful.

It is important to recognize the difference between the behavior of the actual cloud base height and the variation of the ceiling height as it is defined in airway reports. The ceiling usually drops rapidly, especially during the first few hours after the rain begins. However, if the rain is continuous, the true base of the cloud layer descends gradually or steadily. The reason for this is that below the precipitating frontal cloud layer there are usually shallow layers in which the relative humidity is relatively high and which soon become saturated by the rain. The cloud base itself has small random fluctuations in height superimposed on the general trend.

The time of lowering of the ceiling during precipitation remains a local problem. Many excellent sources and references are available for further study. The Navy publication, Synoptic Methods for Developing Forecast Techniques, NWRF 38-0563-073, lists some of the references and gives details for developing a local objective technique.

Time of Lowering of Ceiling

Forecasting the time when the ceiling of a given height will be reached during rain is a separate problem. Nomograms, tables, air trajectories, time air will become saturated, and a correlation of these data can be resolved into an objective technique tempered with empirical knowledge and subjective considerations and can be developed for your individual station. J. J. George and Associates in their studies on Geophysical Research Directorate (GRD) Contracts have developed several studies along these lines.

Extrapolation of Ceiling Trend by Means of the x-t Diagram

The x-t diagram, as explained previously in this chapter, can be used to extrapolate the trend of the ceiling height in rain. The hourly observations should be plotted for stations near a line parallel to the probable movement of the general rain area, originating at your terminal and directed toward the oncoming rain area. Ceiling-time curves for given ceiling heights may be drawn and extrapolated. There may be systematic geographical differences in ceiling between stations due to local (topographic) influences which cause irregularities in the curves. Such differences sometimes can be anticipated from climatological studies, experience, or general influence. In addition, there may be a diurnal (thermal) ceiling fluctuation which will become evident in the curve in slow moving situations. Rapid and erratic up-and-down fluctuations also must be dealt with where the ceiling is uneven due to scud or "holes" of small diameter. In this case, a smoothing of the curves may be necessary before extrapolation can be made. A slightly less accurate forecast may result from this process.

In view of the previous discussion of the precipitation ceiling problem it is not expected that mere extrapolation can be wholly satisfactory at a station when the ceiling lowers rapidly during the first hours of rain, as new cloud layers form beneath the front. However, by following the ceiling trend at surrounding stations as well, a pattern of abrupt ceiling changes may be noted. These changes at nearby stations where rain started earlier may give a clue to a likely sequence at your terminal.

THE TREND CHART AS AN EXTRAPOLATION AID

The trend chart can be a valuable forecasting tool when it is used as a chronological portrayal of a group of related factors. It has the added advantage of helping the forecaster to become "current" when coming on duty. At a glance, the relieving duty forecaster is able to get the picture of what has been occurring at his terminal. Also, he is able to see the progressive effect of the synoptic situation on the weather at his terminal when the trend chart is used with the current map.

Trend Chart Format

The format of a trend chart should be a function of what is desired from it; consequently, it may vary in form from station to station. It should, however, contain those elements which are predictive in nature as well as the quantitative values of parameters to be forecast-such as ceiling and visibility. With this in mind, a suggested format is outlined below.

The trend chart is merely a method for portraying graphically what forecasters generally attempt to store in their memory. Included in the usual technique storehouse of most forecasters is a list of key predictor stations. The forecaster utilizes the hourly and special reports from these stations as aids in making short forecasts for his own terminal. Usually, the sequences from these predictor stations are scanned and committed to memory. Stepwise, the method would be as follows:

1. Determine the direction source of the weather—usually upstream.

2. Select a predictor station(s) upstream and watch for the onset of the critical factor; for example, rain.

3. Note the effect of this factor on ceiling and visibility at predictor station(s).
4. Extrapolate the approach of the factor to determine its onset at your terminal.
5. Consider the effect of the factor at predictor station(s) in forecasting its effect at your terminal.

The chief weakness of this procedure is its subjectivity. The forecaster is required to mentally evaluate all of the information available on the hourly sequences, both for his own and his predictor station(s). The trend chart is a means to graphically portray the sequences and make extrapolations more objective.

The question naturally arises, How many trend charts do I need? The answer depends on the synoptic situation. There are times when keeping a graphic record at all is unnecessary; whereas, at other times the trend for the local terminal may suffice. There should be some blank charts available which may be used to start recording at any time as the need arises. The trend chart format (fig. 10-13) is but one suggested way of portraying the weather record. Experimentation and improvisation are encouraged to find the best form for any particular location or problem.

TIME-LINER AS AN EXTRAPOLATION AID

In the preceding sections of this chapter, several methods have been described for "keeping track of the weather" on a short term basis. Explanations of time-distance charts, isochrone devices, trend charts, etc., have been presented. It is usually not necessary to utilize all or even most of these ideas simultaneously. The device described in this section is designed for use in combination with one or several of the methods previously described. Time-liners are especially useful for isochrone analysis and extrapolation therefrom.

Inasmuch as a large majority of incorrect short range terminal forecasts result from poor timing of weather already occurring upstream, a device such as described below may improve this timing.

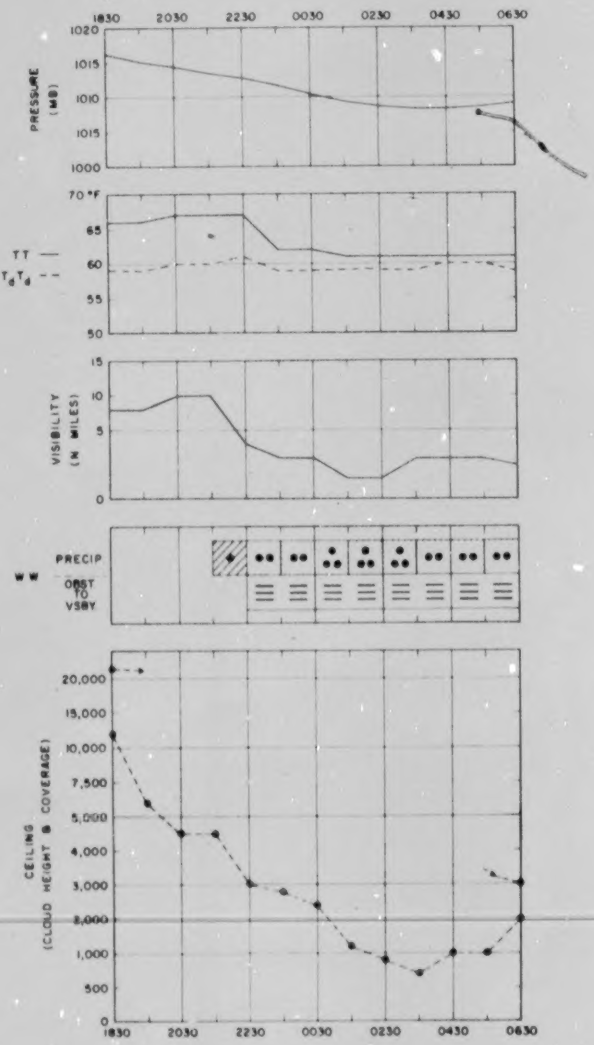

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Figure 10-13.—Trend chart suggested format.

Construction of the Time-Liner

The time-liner is simply a local area map which is covered with transparent plastic and constructed as follows:

1. Using a large-scale map of the local area, construct a series of concentric circles centered on your station and equally spaced from 10 to

20 miles apart. This distance from the center to the outer circle depends on your location, but in most cases 100 to 150 miles is sufficient.

2. Make small numbered or lettered station circles for stations located at varying distances and directions from your terminal. Stations with a high predictor value, relative to your terminal, should be selected. This may be determined by experience, local forecast studies, and climatology. Usually, however, the reporting network is not so dense, and most nearby stations can be spotted. In addition to the station circle indicators, significant topographical features such as rivers, mountains, etc., may be indicated on the base diagram. (Aeronautical charts include these features.)

3. Cover and bind the map with transparent plastic.

Plotting and Analysis of the Time-Liner

By inspection of the latest surface weather map, sequences, and other information, you can determine a section of the diagram and the parameters to be plotted. This will usually comprise about one-third of the circle in a direction from which the weather is approaching. Then, plot the hourly weather SPECIALS for those stations of interest. Make sure to plot the time of each special observation.

Overlay the circular diagram with another piece of transparent plastic and construct isochrones of the parameter being forecast; for example, the time of arrival of the leading or trailing edge of a cloud or precipitation shield. The spacing between isochrones can then be extrapolated to construct "forecast isochrones" for predicting the time of arrival cf occurrence of the parameter at your terminal. Refer to figure 10-14 for an example.

The time-liner is admittedly a "quick fix" tool; however, the fact that it can be plotted and analyzed quickly is a strong point in its favor. If convenient, the distance scale can be constructed to coincide with the scale of the local sectional map used and the isochrones overlaid on the latest analyzed map. This makes a particularly effective briefing aid, since the pilot being briefed can see what is being told by the forecaster regarding the local area. The prime use of this device is, of course, for timing or extrapolation purposes. It cannot be emphasized too strongly that accurate timing is a most important factor in short range terminal forecasting.

USE OF RADAR IN CLOUD AND PRECIPITATION FORECASTING

Radar is highly useful in determining the various clouds that are approaching the station and for estimating the probability of precipitation reaching the local area. Refer to chapter 16 of this training manual for information on forecasting weather conditions utilizing weather-radar.

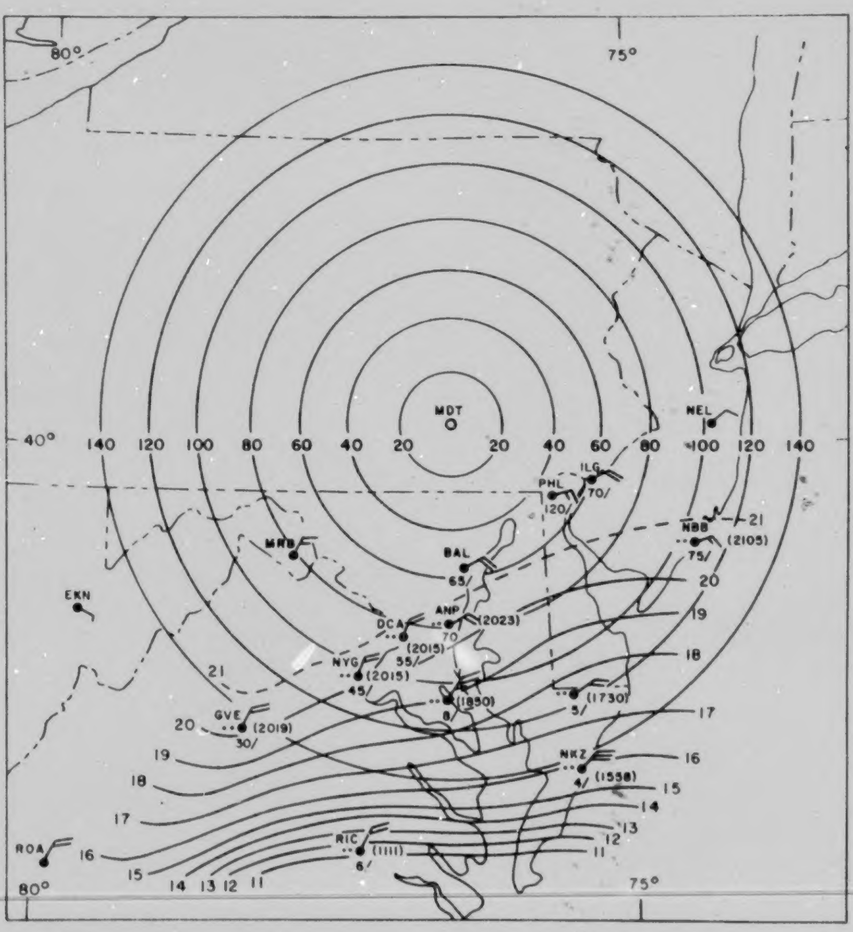
CLOUD ANALYSIS AND FORECASTING

Aerographer's Mates are frequently called upon to make forecasts of clouds for flight and other weather briefings over areas where synoptic observations are not readily available or over other areas where clouds above the lowest layer are frequently obscured by the lower cloud deck. This section is designed to acquaint the AG with the principles of detection and analysis of clouds from radiosonde data. A complete coverage of this problem is beyond the scope of this training manual. Further information on this subject may be found in the following Air Weather Service publications: Forecasting for Aerial Refueling Operations at Mid-Tropospheric Altitudes, AWSM 105-52; and Use of the Skew T, Log P Diagram in Analysis and Forecasting, AWSM 105-124.

IMPORTANCE OF RAOBS IN CLOUD ANALYSIS

The cloud observations regularly available to forecasters in surface synoptic reports leave much to be desired as a basis for cloud forecasting.

Radiosondes which penetrate cloud systems reflect to some extent (primarily in the humidity trace) the vertical distribution of clouds. If the humidity element were perfect, there would usually be no difficulties in locating cloud layers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(2105)
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Figure 10-14.—Large-scale sample time-liner (isochrones show advance of precipitation field).

penetrated by the instrument. Because of the shortcomings in the instrument, however, the relationship between indicated humidity and cloud is far from definite, and an empirical interpretation is necessary. Nevertheless, radiosonde reports give valuable evidence which, when sifted with other available information, aids greatly in determining a coherent picture at least of stratiform and frontal cloud distributions. Their value in judging air mass cumulus and cumulonimbus distribution is negligible.

INFERRING CLOUDS FROM RAOBS

Theoretically we should be able to infer from radiosonde observations of humidity the layers where the sonde penetrated cloud layers. In practice, the determination that can be made from temperature and dewpoint curves are often less exact and less reliable than desired. Nevertheless, raobs give clues about cloud distribution and potential areas of cloud formation. These clues generally cannot be obtained from any other source. A knowledge of the characteristics of the humidity element will aid the forecaster when using data from the sounding.

CHARACTERISTICS OF THE RADIOSONDE HUMIDITY ELEMENT

The carbon-impregnated plastic humidity element is presently used in U.S. radiosondes.

The characteristics of this element are as follow:

1. The time lag is not sensitive to temperature changes, and the polarization effect is negligible.
2. There is no washout problem, although there are some adverse effects due to precipitation actually wetting the element.
3. It will provide good humidity data down to -40°C.
4. The sensitivity is only fair at humidities less than about 15 percent, and there is some uncertainty in the readings near saturation.

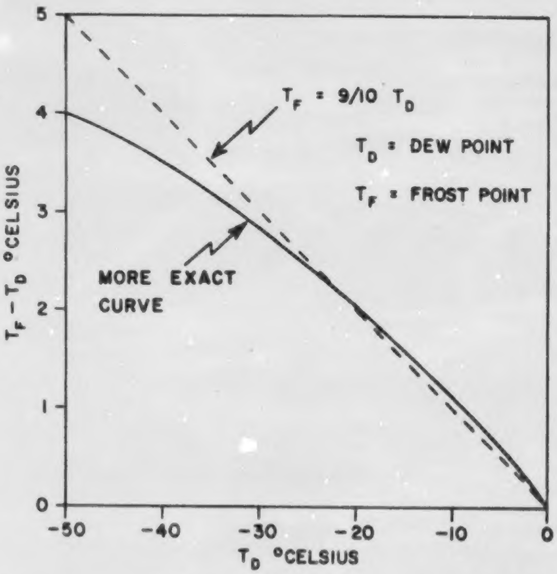

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Figure 10-15.—Difference between frost point and dewpoint as a function of the dewpoint.

DEWPOINT AND FROST POINT IN CLOUDS

The data trace recorded from the humidity strip is calibrated in terms of relative humidity with respect to water at negative as well as positive temperatures. The temperature minus the dewpoint depression value, the humidity parameter that is transmitted over the teletype, gives the dewpoint which is defined as the temperature to which the air must be cooled adiabatically at constant vapor pressure if saturation with respect to a water surface is to be reached. The FROST POINT (that is, the temperature to which the air has to be cooled or heated adiabatically in order to reach saturation with respect to ice) is higher than the dewpoint except at 0°C, where the two coincide. In the graph shown in figure 10-15 the difference between dewpoint and frost point is plotted as a function of the dewpoint itself.

In a cloud in which the temperature is above freezing, the true dewpoint will coincide closely with the true temperature, indicating that the air between the cloud droplets is practically saturated with respect to the water surface of the droplets. Minor discrepancies may occur when the cloud is not in a state of equilibrium (when the cloud is dissolving or forming rapidly, or when precipitation is falling through the cloud with raindrops of slightly different temperature than the air); but these discrepancies are theoretically small. In the subfreezing part of a cloud, the true temperature is between the true dewpoint and the true frost point, depending on the ratio between the quantities of frozen and liquid cloud particles. If the cloud consists entirely of supercooled water droplets, the true temperature and the true dewpoint will, more or less, coincide. If the cloud consists entirely of ice, the temperature should coincide with the frost point. We cannot, therefore, look for the coincidence of dewpoint and temperatures as a criterion for clouds at subfreezing temperatures even if the humidity element has no systematic errors as previously discussed. At temperatures below -12°C, the temperature is more likely to coincide with the frost point than the dewpoint.

The graph shown in figure 10-15 indicates that the difference between the dewpoint and frost point increases roughly 1°C for every 10°C that the dewpoint is below freezing. For example, when the dewpoint is −10°C, the frost point equals −9°C; when the dewpoint is −20°C, the frost point is −18°C; and when the dewpoint is −30°C, the frost point is −27°C. Thus, for a cirrus cloud that is in equilibrium (saturated with respect to ice) at a (frost point) temperature of −40°C, the correct dewpoint would be −44°C (to the nearest whole degree).

We can state then, in general, that air in a cloud at temperatures below about −12°C is saturated with respect to ice, and that as the temperature of the cloud decreases (with height), the true frost point, dewpoint difference increases. The effects of these physical characteristics serve to reinforce the effects of the systematic faults of the humidity element itself, and they all conspire to make a radiosonde ascending through such a cloud indicate increasing dewpoint depressions. Any attempt to determine the height of cloud layers from humidity data of a raob is, therefore, subject to errors from these defects. It is possible to overcome some of these errors by a subjective interpretation of the raobs, as discussed in the following sections.

INTERPRETATION OF RAOBS WITH RESPECT TO CLOUD LAYERS

The following diagrams (figs. 10-16, 10-17, and 10-18) made over the United States illustrate the behavior of the radiosonde during cloud penetration. These are but three of a series of eleven diagrams actually correlated with aircraft observations (from Project Cloud-Trail Flights) or the heights of cloud bases and tops from aircraft flying in the vicinity of the ascending radiosondes. The difference in time and space between the aircraft and sounding observations was usually less than 2 hours and 30 miles. Some of the aircraft reported only the cloud observed above 15,000 feet; others reported all clouds. In figures 10-16 through 10-18, the aircraft cloud observations are entered in the lower left corner of each diagram under the heading cloud; the surface weather report is entered under the aircraft cloud report. Where the low cloud was not reported by the aircraft, the height of the cloud base may be obtained from the surface reports. Aircraft height reports are in pressure-altitude. The temperature, frost point, and dewpoint curves are indicated by T, T_f, and T_d, respectively.

In figure 10-16, a marked warm front is approaching from the south. Moderate continuous rain fell 2 hours later. At 1830Z an aircraft reported solid clouds from 1,000 to 44,000 feet (tropopause). The 1500Z sounding shows an increasing dewpoint depression with height and no discontinuity at the reported cloud top of 15,000 feet. A definite dry layer is indicated between 18,300 and 20,000 feet. The second reported cloud layer is indicated by a decrease in dewpoint depression, but the humidity element

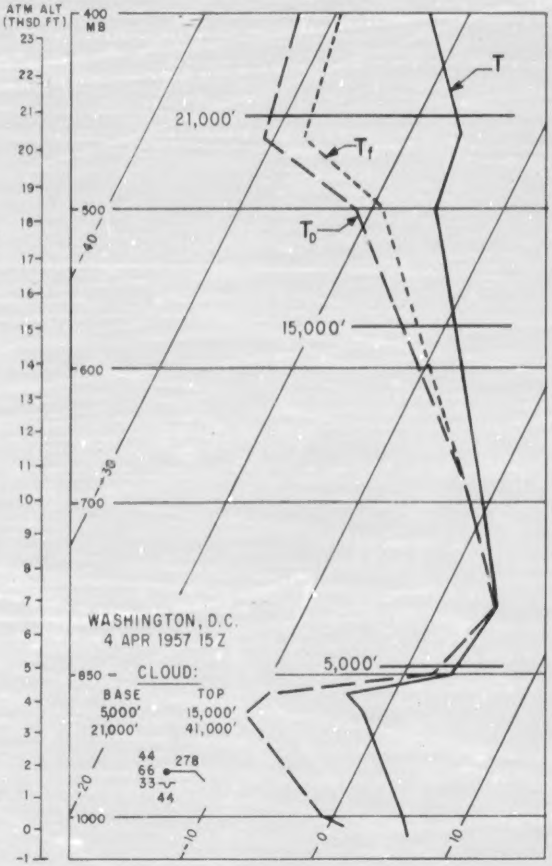

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Figure 10-16.—Example of inferring clouds from a raob with an active warm front approaching from the south.

is obviously slow in responding. The dewpoint depression at the base of the cloud at 21,000 feet is 14°C and at 400 mb; after about a 3-minute climb through the cloud, it is still 10°C. From the sounding, clouds should have been inferred to be from about 4,500 feet (base of the rapid humidity increase) to 500 mb and a second layer from 20,000 feet up. In view of the rapid filling of the cloud free gap between 15,000 and 21,000 feet which followed as the warm front approached, the agreement between reported and inferred conditions is good.

Figure 10-17 shows a middle cloud layer with no precipitation reaching the surface. This is a case of a cloud in the 500-mb surface with no precipitation reaching the surface; the nearest rain was in Tennessee. The evidence from the sounding for placing the cloud base at 12,200 feet is strong, yet the base is inexplicably reported at 15,700 feet. The reported cloud base of 15,700 feet was probably not representative, since altostratus, with bases 11,000 to 14,000 feet, was reported for most stations over Ohio and West Virginia.

Figure 10-18 shows layer clouds with their intermediate clear layers not showing in the humidity trace.

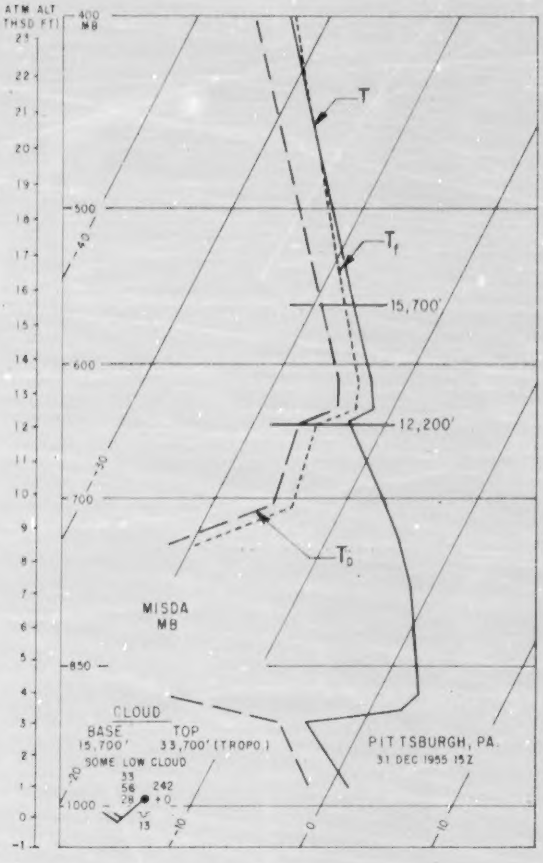

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Figure 10-17.—Example of inferring clouds from a raob with a middle layer and no precipitation reaching the surface.

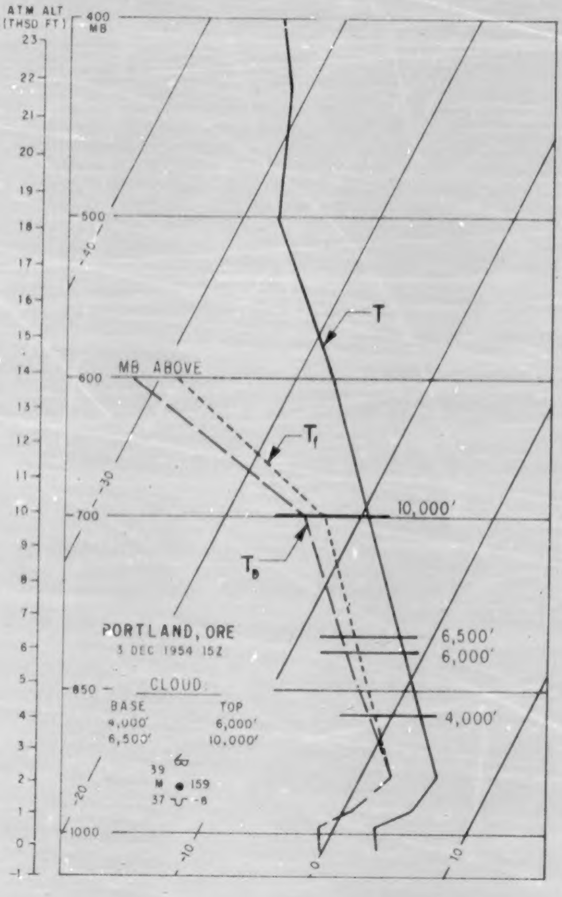

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Figure 10-18.—Example of inferring clouds from a raob, showing layer clouds with their intermediate clear layers not showing in the humidity trace.

humidity trace. There is good agreement between the sounding and the aircraft report. The clear layer between 6,000 and 6,500 feet is not indicated on the sounding. Thin clear layers as well as thin cloud layers usually cannot be recognized on the humidity trace.

Comparisons of the type made in the foregoing between soundings and cloud reports provide us with the following rules:

1. A cloud base is almost always found in a layer (indicated by the sounding) where the dewpoint depression decreases.
2. The dewpoint depression usually decreases to between 0°C and 6°C when a cloud is associated with the decrease. In other words, you should not always associate a cloud with a layer of dewpoint decrease, but only when the decrease leads to minimum dewpoint depressions from 6°C to 0°C; at cold temperatures (below -25°C), however, dewpoint depressions in clouds are often higher than 6°C.
3. The dewpoint depression in a cloud is, on the average, smaller for higher temperatures. Typical dewpoint depressions are 1°C to 2°C at temperatures of 0°C and above, and 4°C between -10°C and -20°C.
4. The base of a cloud should be located at the base of the layer of decreasing dewpoint depression, if the decrease is sharp.
5. If a layer of decrease of dewpoint depression is followed by a layer of stronger decrease, the cloud base should be identified with the base of the strongest decrease.
6. The top of a cloud layer is usually indicated by an increase in dewpoint depression. Once a cloud base is determined, the cloud is extended up to a level where a significant increase in dewpoint depression starts. The gradual increase of dewpoint depression with height that occurs on the average in a cloud is not significant.

In addition to the above analysis of Project Cloud-Trail data, another study was made by the U.S. Air Force to see how reliable the dewpoint depression is as an indicator of clouds. The results of this study are summarized in the two graphs shown in figure 10-19. Each graph shows the percent probability of the existence of a cloud layer in January for different values of dewpoint depression. On each graph one curve shows the probability of clear or scattered conditions as a function of the dewpoint depression; the other curve shows that of broken or overcast conditions. Separate graphs are included for the 1,000- to 850-mb and 850- to 600-mb layers. The graphs are based on 1,027 observations, which are enough to indicate the order of magnitude of the dewpoint depressions at the base of winter cloud layers. Minor irregularities in the curves were not smoothed out because it is not certain that they are all due to insufficient data. The graphs are applicable without reference to the synoptic situation. For a given winter sounding, one can estimate from

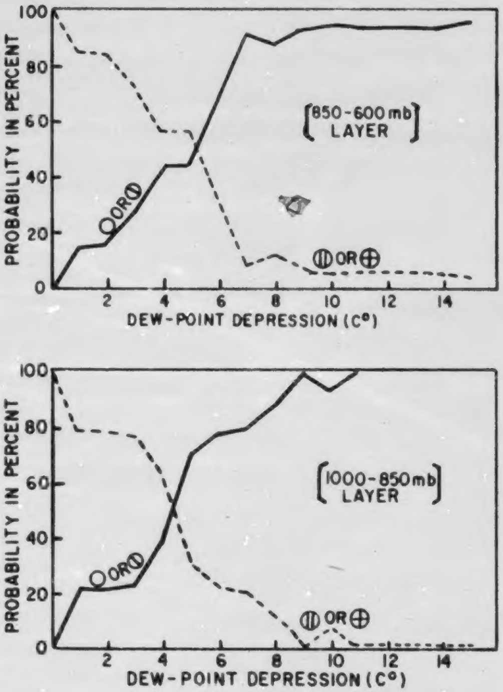

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Figure 10-19.—Percent probability of existence of cloud layer bases for different values of dewpoint depression (degrees C). Solid lines represent probability of clear or scattered conditions; dashed lines, the probability of broken or overcast conditions with the cloud layer bases between 1,000 mb and 600 mb.

the graph the probability of different sky cover conditions with cloud bases between 1,000 mb and 600 mb for layers of given minimum dewpoint depressions.

HUMIDITY FIELD IN THE VICINITY OF FRONTAL SYSTEMS

Studies of the humidity field in the vicinity of frontal zones indicate there is a tongue of dry air extending downward in the vicinity of the front and tilted in the same direction as the front. One study found that such a dry tongue was more or less well developed for all frontal zones investigated. This dry tongue was best developed near warm fronts; it extended, on the average, down to 700 mb in cold fronts and to 800 mb in warm fronts. In about half the number of fronts, the driest air as found within the frontal zone itself; on occasions it was found on both the cold and warm sides of the zone. About half the flights through this area in connection with this study showed a sharp transition from moist to dry air, and the change in frost point (to be explained later in this chapter) on these flights averaged about 20°C in 35 miles. Some flights gave changes of more than 20°C in 20 miles.

As a frontal cloud deck is approached, the dewpoint depression (or frost point depression) starts diminishing rapidly in the close vicinity of the cloud. Farther away then 10 to 15 miles from the cloud, the variation was much less and was less systematic. This fact should be borne in mind when attempting to locate the edge of a cloud deck from raob humidity data (500 mb, for instance). Linear extrapolation or interpolation of dewpoint depressions cannot be expected to yield good results. For instance, when one station shows a dewpoint depression of 10°C and the neighboring station shows saturation, the frontal cloud may be anywhere between them, except within about 10 miles of the driest station.

Since frontal cloud masses at midtropospheric levels are usually surrounded by relatively dry air, it is possible to locate the edge of the cloud mass from humidity data on constant pressure charts, at least to within the distance between neighboring soundings, even if the humidity data are not very accurate. This is so because the typical change of dewpoint depression in going from the cloud edge, a distance of 10 to 15 miles or more, into cloud free air is considerably greater than the average error in the reported dewpoint depression.

500-MB ANALYSIS OF DEWPOINT DEPRESSION

Figure 10-20 shows an analysis of the 500-mb dewpoint depression field, superimposed upon an analysis (based on surface observations) of areas of continuous precipitation and of areas of overcast middle clouds. The 500-mb dewpoint depression isopleths were drawn independently of the surface data. The analysis shows that:

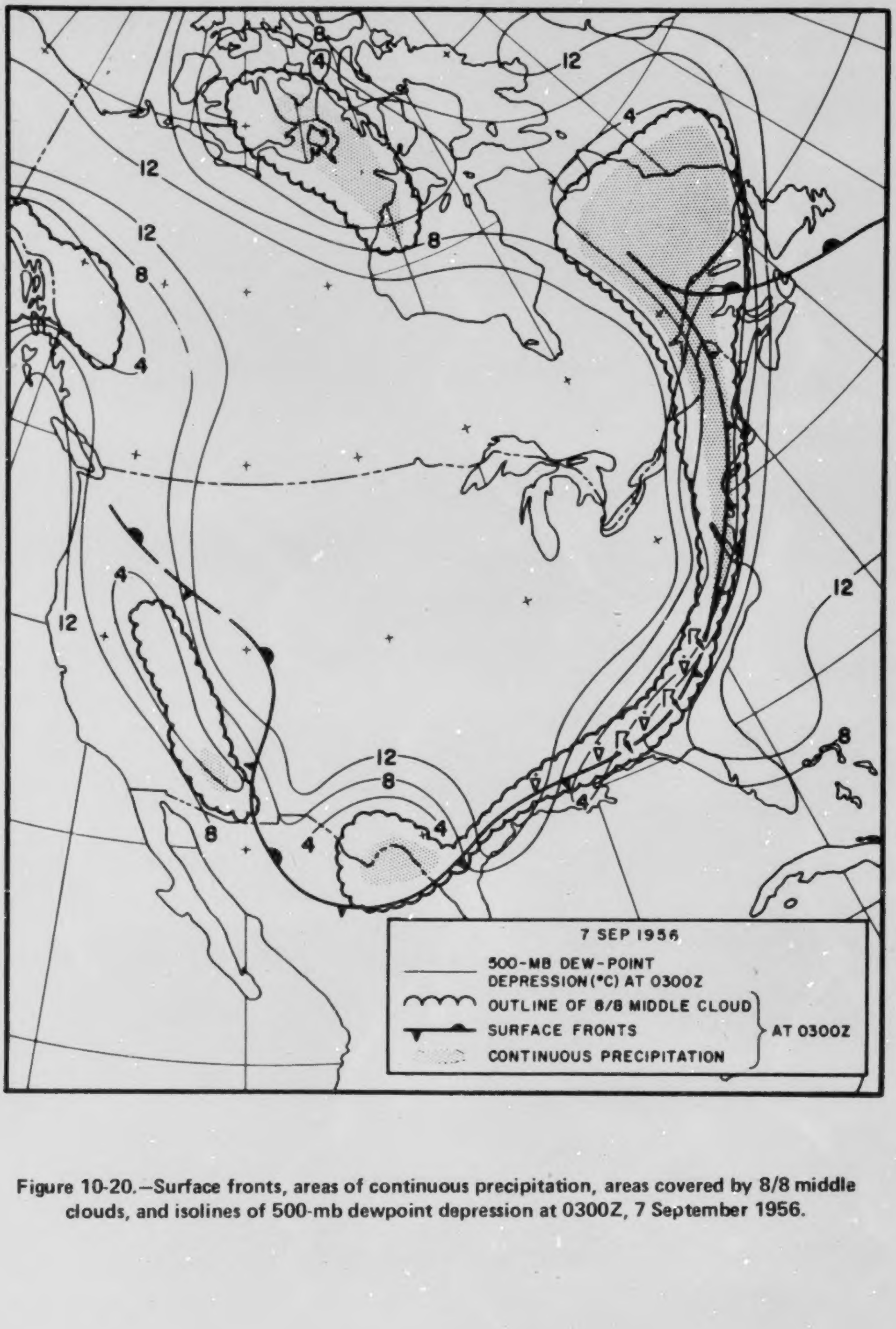
1. The regions of high humidity at 500 mb coincide well with the areas of middle cloud and the areas of precipitation.
2. The regions of high humidity at 500 mb are separated from the extensive dry regions by strong humidity gradients. These gradients are, in all probability, much stronger than shown on this analysis, since this analysis has the defect of all continuous field analyses which are based on discrete observations spaced widely apart; in other words, linear interpolation between observations smoothes out strong contrasts.
3. A dewpoint depression of 4°C or less is characteristic of the larger part of the areas of continuous precipitation and also of the larger part of the area of overcast middle clouds.

Since the 500-mb dewpoint depression analysis agrees well with the surface analysis of middle cloud and precipitation, the possibility exists of replacing or supplementing one of these analyses with the other.

The characteristics of the 500-mb dewpoint depression analysis (outlined above) make it a valuable adjunct to the surface analysis. These analyses can be compared and, by crosschecking, each can be completed with greater accuracy than if they were done independently. A rough sketch of the dewpoint depression field may be completed on the facsimile chart in a matter of minutes.

THREE-DIMENSIONAL HUMIDITY ANALYSIS–THE MOIST LAYER

A single level (for example the 500-mb level) dewpoint depression analysis to find probable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7 SEP 1956
500-MB DEW-POINT DEPRESSION (°C) AT 0300Z
OUTLINE OF 8/8 MIDDLE CLOUD
SURFACE FRONTS
CONTINUOUS PRECIPITATION
AT 0300Z

AG.594

Figure 10-20.—Surface fronts, areas of continuous precipitation, areas covered by 8/8 middle clouds, and isolines of 500-mb dewpoint depression at 0300Z, 7 September 1956.

cloud areas does not indicate clouds above or below that level. For example, if the top of a cloud system reached only to 16,000 feet, and there was dry air above at 500 mb, you would never suspect, from the 500-mb analysis, the existence of clouds a short distance below the 500-mb level.

However, an analysis of the extension of the moist layers in three dimensions can be obtained simply by scrutinizing each individual raob. The raobs selected should be those for the general vicinity of, and the area 500 to 1,200 miles upstream of, the area of interest, depending on the forecast period. (Most systems during a 36-hour period will move less than 1,200 miles.) The moist layers apart from the surface layers can be determined by methods previously discussed in this chapter. The heights of the bases and tops (labeled in thousands of feet of layer is present, this fact can be indicated, though there is little advantage in indicating a dry layer 2,000 to 3,000 feet (or less) thick sandwiched between thicker moist layers. Usually, it is sufficient to indicate the entire moist layer, without bothering about any finer structure. A survey of the cloud field is made easier by writing the heights of the bases and tops in different colors.

A moist layer for the sake of simplicity may be defined as a layer having a frost point depression of 3°C or less (i.e., a dewpoint depression of 4°C at −10°C; 5°C at −20°C; 6°C at −30°C).

LIMITATIONS TO DIAGNOSIS OF TOWERING CUMULUS AND CUMULONIMBUS DISTRIBUTION FROM RAOBS

The cumulus and cumulonimbus of summer and tropical air mass situations are generally scattered. In many, if not most cases, they do not actually cover half of the sky. Under such conditions the probability of a radiosonde, released one to four times daily at a fixed time and place, passing up through a cloud of this type appears to be small. When a balloon does enter the base of a tall cumulus cloud, it is likely to pass out of the side of the cloud rather than the top, or it may get caught for a time in a downdraft, giving an ambiguous record of vertical cloud distribution.

For the above reasons, experience indicates that little dependence can be placed on the usual sounding to indicate directly the existence of tall cumulus in the area. On the other hand, where the radiosonde samples of the environment of such clouds, a stability analysis, combined with considerations of surface weather observations, radar and aircraft reports, and synoptic analysis for heating and convergence, usually provides an estimate to the extent of cumulus sky coverage.. This approach uses the same principles as in thunderstorm and severe weather forecasting. (See chapter 11 of this training manual.)

In those cases where the sounding passes up through a cumulus, it is well to keep in mind that the temperature in parts of such clouds is often colder than the environment just outside the clouds. These colder regions may still have buoyancy relative to other parts of the cloud surrounding them. Also, old dissipating cumulonimbus clouds are generally colder than their environment.

The lapse rate in cumulus and cumulonimbus is not necessarily saturation adiabatic due to the effects of "holes," downdrafts, melting of snow or hail, entrainment, mixing, etc.

PRECIPITATION AND CLOUDS

The type and intensity of precipitation observed at the surface is related to the thickness of the cloud aloft, and particularly to the temperatures in the upper part of the cloud. The processes which cause cloud particles to grow and precipitate out of clouds have received much attention in the last decade. Our knowledge of these processes is far from complete, but much information of practical use to the forecaster is available.

The results of one study relating cloud-top temperatures to precipitation type and intensity are given as follows: From aircraft ascents through stratiform cloud and correlated surface observations of precipitation, it was found that in 87 percent of the cases when drizzle occurred, it fell from clouds whose cloud-top temperatures were higher than −5°C; the frequency of rain or snow increased markedly when the cloud-top temperature fell below −12°C. When continuous rain or snow fell, the temperature of the coldest part of the cloud was below -12°C in 95 percent of the cases. Intermittent rain was mostly associated with cold cloud-top temperatures. When intermittent rain was reported at the ground, the cloud-top temperature was below -12°C in 81 percent of the cases and below -20°C in 63 percent of the cases. From this, it appears that when rain or snow—continuous or intermittent—reaches the ground from stratiform clouds, the clouds—solid or layered—extend in most cases to heights where temperature is well below -12°C, or even -20°C.

This rule cannot be reversed. When no rain or snow is observed at the ground, middle clouds may well be present in regions where the temperature is below -12°C or -20°C. Whether or not precipitation reaches the ground will depend on the cloud thickness, height of the cloud base, and the dryness of the air below the base.

INDICATIONS OF CIRRUS CLOUDS IN RAOBS

True cirrus forms at temperatures near -40°C or colder and consists of ice crystals. At these temperatures, as soon as the air is brought to saturation with respect to water, the condensate immediately freezes. The crystals then often descend slowly to levels of -30°C, and persist for a long time if the humidity below the formation level is high enough. In general, cirrus is found in layers which have saturation or supersaturation with respect to ice (at any temperature colder than 0°C; if the relative humidity with respect to water is 100 percent, then the relative humidity with respect to ice is greater than 100 percent).

Present radiosonde humidity elements are far from capable of measuring humidity values satisfactorily at the temperatures of the cirrus levels. However, the elements will often show a change in humidity at low temperature that reflect the presence of moist layers which contain cirrus.

Several studies have pointed to indirect indications of cirrus presence whenever the dewpoint depression at 500, 540, and 400 mb was relatively low.

A study published by the U.S. Air Force has shown how contrail forecasting curves can be used to improve the accuracy of ground observer estimates of the height of observed cirrus layers. The method is published in AWS TR 105-110. Memorandum on Estimating the Height of Cirrostratus Clouds, and a test with comments on the method was published in AWS TR-110A, Test of Method for Estimating the Height of Cirrostratus Clouds.

CIRRUS FORECASTING

One of the senior Aerographer's Mates duties is to make forecasts of cirrus or cirrostratus coverages and heights. This poses difficulties owing to the lack of proven techniques of universal applicability and even more to inadequacies of synoptic observations and of upper air soundings. For a complete review of the available knowledge of high clouds and several forecasting methods, it is suggested you thoroughly study the Air Weather Service publication. A Compendium on Cirrus and Cirrus Forecasting, AWS TR 105-130. This publication contains a comprehensive study of the nature of cirrus, the physics of cirrus formation, cirrus in relation to certain phenomena, cirrus climatology, cirrus types, and a review of several methods of forecasting cirrus with comments regarding their limitations and/or value. Also, refer to chapter 4, Aerographer's Mate 3 & 2, NavTra 10363-D, and chapter 3 of this training manual for additional background of cirrus types and the physics of formation.

OBSERVATION AND FORMATION OF CIRRUS

Cirrus (or "cirriform") clouds are conventionally divided into three general groups: cirrus (proper), cirrostratus, and cirrocumulus. The latter is a relatively rare cloud and may be ignored for practical purposes. Cirrus proper (detached or patchy cirrus) does not usually create a serious operational problem. Cirrostratus and extensive cirrus haze, however, are troublesome in high level jet operations, for aerial photography, interception, rocket tracking, and guided missile navigational systems. Therefore, a definite requirement for cirrus forecasting exists. No universally applicable nor completely successful forecasting method has been found, though several methods have been advanced which seem to have some promise of success. One such method is presented in this chapter.

The initial formation of cirrus normally requires that cooling take place to saturation with respect to water and to temperatures near -40°C. Under these conditions, water droplets are first formed but most of them immediately freeze. The resulting crystals persist as long as the environmental humidity remains near ice saturation;, a deep ice-saturated layer usually exists just below the cirrus formation level, which permits long cirrus streamers to descend (slowly) to lower levels and to persist for many hours (or even perhaps days) before evaporating. There is some evidence that the speed of the cooling and the kind and abundance of freezing nuclei may have an important effect on the form and occurrence of cirrus. Slow ascent (cooling) starts crystalization on specially favorable freezing nuclei at humidities substantially below saturation with respect to water; this is presumably the case in extensive cirrostratus associated with warm front altostratus. If slow ascent occurs in air having insufficient freezing nuclei, a widespread haze may result which even at -30° to -40°C is predominantly of water drops. In the case of more rapid cooling (ascent), there is a tendency for the first condensation to contain a higher proportion of water drops, leading to a "mixed cloud" (ice and water) which will convert to ice or snow in time. Presumably, dense cirrus, fine cirrus, cirrocumulus, and anvil cirrus are of this rapid ascent type; it has been postulated, for example, that fine cirrus (proper) is formed in shallow layers undergoing rapid convection due to advection of colder air at top of shear layer.

On the other hand, the fine cirrus and the cirrostratus are so often associated, and cirrostratus is so often reported by pilots as developing from the merging of fine cirrus streamers, that there is question whether the process of formation in cirrus and cirrostratus is essentially different. Nevertheless, the prevailing crystal types in cirrus and cirrostratus seem to differ, though this may not be universal or may merely represent different ages in a characteristic cirrus evolution.

Horizontal visibilities within extensive cirrostratus over middle latitudes are generally between 500 feet and 2 miles. But even thin cirrus haze, invisible from the ground, often reduces the visibility to 3 miles. Burton's empirical rule for forecasting the visibility has been successful for the Arctic cirrus and may work elsewhere.

Captain M. W. Burton of the Air Weather Service devised a rule of thumb for forecasting or estimating the visibility of one aircraft from another in thin cirrus or other high cloud (temperatures below -30°C). This rule is: Visibility = 1/2 mile times dewpoint depression in degrees C. For example: Temperature is -35°C, dewpoint is -38°C, then visibility = 1/2 X 3 = 1 1/2 miles. This rule has been used successfully in the Arctic where poor visibilities in apparently cloud free air are often encountered. Its applicability elsewhere remains to be tested.

The climatology of cirrus occurrence is unsatisfactory because the usual ground observations (in humid climates) miss more than 50 percent of the true frequency or amount, and aircraft observations are very scarce.

Radar cloud detection sets appear to promise much more complete cirrus observations than possible from visual observers.

A cirrus forecasting method, whether empirical or with a physical hypothesis, should exploit what is already known about cirrus in relation to other forecastable parameters. A considerable number of such parameters have been tried in the course of past studies, in addition to pure extrapolation of the movement of cirrus cloud (neph) systems. Except for the use of the surface pressure pattern, these are all upper air parameters. Most of them were tried or chosen on the generally accepted hypothesis that cirrus, like other clouds, forms where there is sufficient large scale vertical motion and the initial humidity is relatively high.

Most of the various parameters chosen in the more empirical procedures which have been used or proposed for cirrus forecasting can be given a rational interpretation as having indirect relations to humidity or vertical motion. For example, the models of frontal, pressure, contour, and wind patterns associated with cirrus are in this category. Such methods are very subjective and depend on considerable experience for success. But other hypotheses, such as a relation of the temperature, or the lapse rate, or the wind shear, to cirrus occurrence or height, do not seem to find much convincing statistical support nor a clearly defensible physical basis. Statistical techniques for local cirrus forecasting tried thus far have not been too encouraging, but with better data and more experience with other techniques further trials of this sort should be worthwhile.

CIRRUS FORECASTING PROBLEM

Only a limited number of methods of forecasting cirrus type clouds have been developed and published. Obviously, this field is open to further investigation, testing, and study. Many forecasters have attempted to forecast cirrus or high clouds using familiar frontal or cyclone models. None of these methods or models alone have proven operationally adequate, though several of the special methods show some success or appear promising for further development. There are a number of parameters, both surface and aloft, which have been correlated with cirrus occurrence or formation. A few of the more prominent ones are mentioned here.

Surface Pressure Pattern

When weather forecasting was based solely on surface iosbaric analysis, many meteorologists attempted to find an empirical statistical relationship between the directions and speeds of cirrus in advance of depressions and the subsequent weather. With the advent of the Norwegian School of synoptic meteorology, frontal and cyclone models are developed which embodied the associated idealized cloud distribution. In these models the cirrus thickening and lowering into altostratus is a characteristic sequence in an advancing warm front. The cirrus of fair weather outside the cyclone cloud shields is neither identified nor accounted for in the Norwegian models. It is well known that much cirrus is observed which has no obvious relation to any features analyzed on surface synoptic charts.

Fronts Aloft

Above 500 mb the usual concepts of air masses and fronts have little application. Most of the fine cirrus seen ahead of (and above) warm fronts or lows initially forms detached from the frontal middle cloud shield, though later it may trail downward to join the altocumulus and altostratus. One study reveals that with precipitation occurring in advance of a warm front, a 60 percent probability exists that cirrus is occurring above this zone. Cirrus observed with the cold front cloud shield either originates from cumulonimbus along and behind the front or from convergence around an associated upper trough. In many cases there is no post cold front cirrus, probably due to marked subsidence aloft. Squall lines can also produce much anvil cirrus which spreads out in advance and persists after the cumulonimbus dissipates.

Contour or Flow Direction Aloft

Studies in the past relating to motions of cirrus and the ensuing motions of surface lows and highs lead to many useful rules that have been familiar to forecasters for generations.

Data on the flow direction aloft are scanty, and direction of the wind alone appears, by itself, to be a poor parameter. Therefore, direction must be considered along with other parameters at upper levels.

However, one rule which should be of value as revealed by several studies showed that the percentage of high cloud is greatest when the wind at 300 mb is from the southwest to west and least when from the northeast to east. The inverse is true for no high cloud relation. No relation with wind speed was evident.

Contour Patterns Aloft

One of the forecasting rules used widely in the United States for several decades or more states that the ridge line at 20,000 feet (about 500 mbs) preceding a warm front marks the forward edge of the cirrus cloud sheet. This undoubtedly refers to the edge of the solid cirrostratus-altostratus overcast. It probably does not include fine cirrus, which would either be higher and/or would not evaporate immediately in the lee of the ridge.

Commanders Wolff and Somervell of Project AROWA devised a set of rules of this type in which both surface and 500-mb patterns are considered. For a typical 500-mb wave pattern, they state:

1. No extensive cirrostratus will occur before the surface ridge line arrives.
2. Extensive cirrostratus follows the passage of the surface ridge line.
3. No middle clouds appear before the arrival of the 500-mb ridge line.
4. Middle clouds tend to obscure the cirrus after the 500-mb ridge line passes.

When the 500-mb wave is rather flat the cirrus arrival is delayed and the cloud is thinner. The greater the 500-mb streamline convergence from trough to ridge, the more cirrus between the surface and 500-mb ridge lines.

Cirrus in Relation to the Tropopause

Experiences of pilots of high-flying aircraft have confirmed the earlier theory that the tops of most cirrus are at or below the tropopause. In midlatitudes the top of most extensive and thick cirrus layers is at or within several thousand feet of the polar tropopause height; only some patchy cirrus is found between the equivalent polar tropopause height and the tropical (high) tropopause. A small percentage of cirrus cases (including sometimes extensive cirrostratus) is observed in the lower stratosphere above the polar tropopause, up to 50,000 feet, but mainly below the level of the jetstream core. The cirrus of the equatorial zone also generally extends to the tropopause. There is a general tendency for the mean height of the bases to increase from high to low latitudes more or less parallel to the mean tropopause height, ranging from 24,000 feet at 70°-80° latitude to 35,000 to 40,000 feet or higher around the Equator. The thickness of individual cirrus layers (the clouds are often multilayered) is most frequently 800 feet in midlatitudes; some cases, however, range up to 10,000 feet or more with the average cirrus zones about 6,500 feet. The mean thickness of cirrus affected zones tends to increase from high to low latitudes. In polar continental regions in winter, cirrus virtually comes down to the ground. In midlatitudes and in the Tropics, there is little seasonal variation.

Cirrus in Relation to the Jetstream

A discussion of cloud types associated with the jetstream is contained in chapter 4 of this training manual. However, a few additional observations and relations are presented here. All of the studies made in this relation agree that most of the more extensive and dense cirrus is on the high-pressure (right, or south) side of the jet axis. Much of the largely observed frequency of high clouds well north of the jet axis can probably be accounted for as the upper reaches of cold frontal systems or cold lows not directly connected with the jetstream. In some parts of a trough, this high cloud may tend to be dense and in other parts thin or scattered.

A CIRRUS FORECASTING PROCEDURE

The following technique was originally developed by F. Singleton and B.G. Wales-Smith of the British Meteorological Service and was published under the title, A Note on Cirrus Forecasting, in the April 1960 issue of The Meteorological Magazine, No. 1,053, Vol. 89. An abbreviated method of using this procedure was developed by the Air Weather Service, 2nd Weather Wing, Forecasters Bulletin C-13, and was published in Armed Service Technical Information Agency Bulletin, AD 252903. Credit is given to both sources for permission to use the material in the following section.

It must be realized that this is only one among several techniques available. It was selected for its simplicity and ease of application. However, all the other rules and empirical knowledge should be applied simultaneously to obtain the forecast.

British forecasters have used as an objective means of forecasting cirrus, worksheets consisting of 13 questions for making a 6- to 9-hour forecast, and 6 questions for making a 24- to 36-hour forecast. Although worksheets are based on work done at the British Meteorological Office, the parameters used should have universal application in that they are based on generally accepted rules and theories on cirrus forecasting.

A recent investigation has shown that answers to just 3 of these questions not only provide a good basis for a "Cirrus" or "No Cirrus" forecast but also may be used to obtain an indication of the expected high cloud cover.

The technique utilizing the 3 questions is valid as far ahead as a satisfactory synoptic forecast can be made. We will assume that it applies to a forecast period of 24 to 36 hours.

It is the purpose of this modification to combine the worksheets and the technique, utilizing only 3 questions into a single forecasting procedure.

The procedure is covered in the following four steps:

Step A. Answer the 3 questions in Part A of the combined worksheet (figure 10-21). The questions are answered from prognostic charts for the 24- to 36-hour forecasts and by extrapolating from current charts for the 6- to 9-hour forecast. For each question give a "score" of 0, 1/2, or 1, according to whether the answer is "No," "Uncertain," or "Yes." With the score thus obtained, go to Step B.

Step B. Enter figures 10-22 and 10-23 with the score obtained in Step A. Figure 10-23 shows cirrus amounts expressed in cumulative percentage as a function of the score. Figure 10-23 indicates the most likely amount of high cloud to be forecast for a given score. As an example of the use of the figures, a score of 1, obtained in Step A, indicates that the most likely amount of cirrus is 2 oktas while a forecast range of 0-3 oktas will be correct 61 percent of the time. This and other examples are listed in table 10-1.

Examination of table 10-1 shows that for scores of 0-1 and 2 1/2 -3, fairly definitive forecasts are obtained. For these scores, the procedure stops here.

But with a score of 1 1/2 or 2 the forecast is indeterminate. For example, a score of 2 indicates a forecast range of 0-4 oktas is likely on 51 percent of occasions and a range of 2-7 oktas is likely on 48 percent of occasions. In order to decide between these alternatives, we proceed to Step C for a 24- to 36-hour forecast or to Step D for a 6- to 9-hour forecast.

Step C. Answer the questions in Part B of the worksheet. If one or more of the questions can be answered affirmatively, then the greater amount of cirrus is indicated. If not, then a forecast of the lesser amount of cirrus is suggested.

Step D. Answer the questions in Part C of the worksheet. If four or more of the questions can be answered affirmatively, then the greater amount of cirrus is indicated. If not, then a forecast of the lesser amount is suggested.

By applying the three basic questions in the outlined manner and by supplementing the resulting forecast where necessary with the remaining questions of the worksheet, a cirrus forecasting method is obtained which is believed to have considerable merit.

The authors note that the criteria employed are not adequate for forecasting convective cirrus and that no conclusions can be drawn as to the validity of the technique for this purpose.

PREDICTION OF SNOW VS RAIN

The problem of predicting snow vs rain is, however, important in its own right because many decisions of great operational importance may hinge on the forecast. Ordinarily, an inch or so of precipitation in the form of rain will cause no serious inconvenience. On the other hand, the same amount of precipitation in the form of snow (or sleet or freezing rain) can seriously interfere with transportation, communications, and other naval operations. In such cases the snow-rain problem becomes a factor of the greatest operational significance.

This section is concerned chiefly with the forecasting of snow vs rain, assuming that precipitation can be correctly predicted. The intermediate elements, sleet and freezing rain, which often occur in the boundary zone between snow and rain, are generally grouped with snow in most of the treatment here. However, one objective technique explains a method of differentiating between the several types. Specifically this section deals primarily with forecasting techniques which bear directly upon the snow-vs-rain problem in the United States. The relationships, however, should have a worldwide application insofar as the parameters that are considered. The values, or course, would have to be modified in accordance with your geographical location. This should be easily accomplished through a local study of the optimum conditions. The various techniques and systems given

Table 10-1. —Optimum amounts of cirrus to forecast for "score" 0-3 and examples of expected percentage of correct forecasts.

"Score"	Optimum amount (oktas)	Correct forecasts	
		Range in oktas	Percent
0	0	0-1	66
1/2	1	0-2	62
1	2	0-3	61
1 1/2	3	0-3	50
2	4	0-4	51
—	-	2-7	48
2 1/2	5	5-8	53
3	6-7	5-8	59

PART A

(24-36 hour forecast period. Use prognostic charts)

The three questions to be answered initially are:

1. Is the forecast area in or just to the rear of a ridge in the 200-mb (300-mb)* contour pattern?
2. Is the forecast area on the anticyclonic side of a 200-mb (300-mb) jetstream and within 300 miles of the jet axis?
3. Is the forecast area less than 300 miles ahead of a surface warm front or occlusion?

*Questions 1 and 2 in Part A were evaluated with 300-mb data originally but with 200-mb data in the abbreviated method. It is a reasonable assumption that either chart may be used since a ridge position or anticyclonic shear will normally appear at the same location on both charts.

PART B

(24-36 hour forecast period. Use prognostic charts)

1. Is the air likely to be moist? (Using the 500-mb progged contours and latest available actual chart, the air at this level should be traced back, from the area at the time for which the forecast is required, to the region of an available radiosonde ascent. By examining the depression of dewpoint below temperature at the levels 500, 450, and 400 mb, answer the question, Is the dewpoint depression less than or equal to 10° C, at or above 500 mb?).
2. Will the area be in a thermal ridge as suggested by the 1,000-500 mb thickness prog?
3. Will the 300-mb wind over the area veer from the 500-mb wind by 20° or more?

PART C

(6-9 hour forecast period. Extrapolate from current charts)

1. Is the depression of dewpoint below air temperature at 500 mb less than or equal to 10°C?
2. Is the depression of dewpoint below air temperature at 450 mb less than or equal to 10° C?
3. Is the depression of dewpoint below air temperature at 400 mb less than or equal to 10° C?
4. Is the lapse rate in the 500- to 300-mb layer greater than the wet adiabatic?
5. Is the 400-mb wind between SW and NW?
6. Is there a veer of wind between 500 and 300 mb of 20° or more?
7. Is the 1,000-500 mb thermal wind greater than 20 KT?
8. Is the forecast area in a ridge in the 1,000-500 mb thickness pattern?
9. Is there anticyclonic curvature of the 1,000-500 mb thickness lines?
10. Is there a deep cold pool or intense thickness trough in the 1,000-500 mb thickness pattern?

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Figure 10-21.—Cirrus forecasting worksheet.

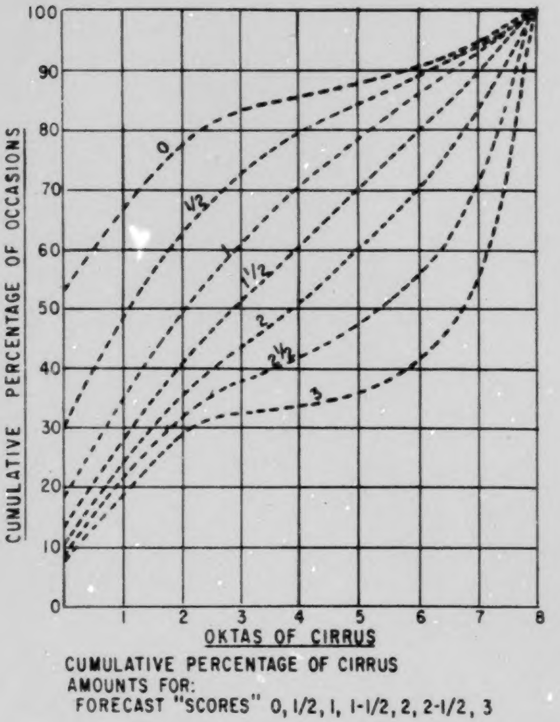

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Figure 10-22.—Cumulative percentage of cirrus amounts for forecast "scores" 0, 1/2, 1, 1 1/2, 2, 2 1/2, 3.

here often complement each other, and the alert forecaster should be aware of the application of all methods. The approach used here is a discussion of the general synoptic patterns, the thermal relationships (that is, the use of temperatures at the surface and aloft in separating cases of rain from those of snow), and the presentation of an objective technique to distinguish the types of precipitation.

The material in this section of the chapter was derived chiefly from two publications: The Prediction of Snow vs Rain, U.S. Department of Commerce Forecasting Guide No. 2; and Further Studies in the Development of Short Range Weather Prediction Techniques, GRD Contract No. AF19(604)-2073, Scientific Report No. 1. A more complete study of these and other publications should immeasurably improve your ability to differentiate between frozen and liquid forms of precipitation in your forecast.

GEOGRAPHICAL AND SEASONAL CONSIDERATIONS

The forecasting problem of snow vs rain arises during the colder portion of the year. However, in midwinter when the problem is most serious in the Northeastern and North Central States, some other regions of the United States are not at all concerned. For example, in the Rocky Mountain region snow is nearly the exclusive type of precipitation in midwinter. Along the Gulf of Mexico and the Pacific coasts (excluding the Sierra Nevada and Cascade Ranges), snow is rare. Other areas of the world experience similar problems.

A set of figures has been reproduced in the U.S. Department of Commerce's Snow vs Rain publication which demonstrates the geographic and seasonal variations of snow vs rain. These figures show the normal monthly snowfall in inches and the total melted precipitation in inches, as well as the ratio between snowfall and total precipitation. These data were obtained

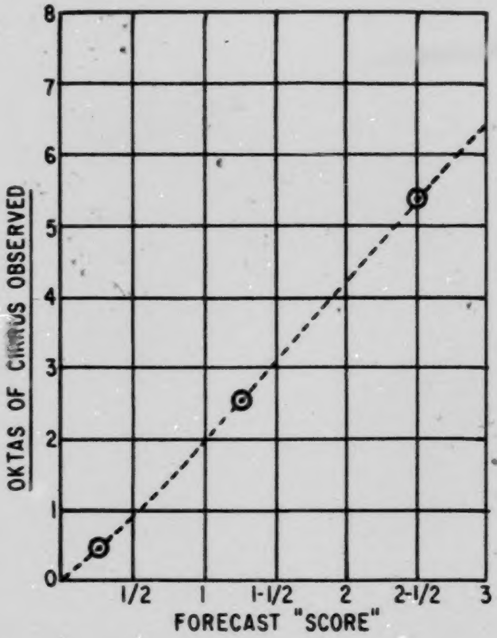

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Figure 10-23.—Average amounts of cirrus observed plotted against forecast "score."

from the National Weather Service. Local Climatological Data for various stations. It is assumed for convenience that on the average 1 inch of snow yields about 1/10 inch of melted precipitation, then an "all-snow" station would have a ratio of 10 a station with 50 percent of its precipitation as snow a ratio of 5, and a "no snow" station a ratio of 0. It should be mentioned that recent studies indicate the ratio of snowfall to its water equivalent may average as much as 12 or 13, even when temperatures are not far below freezing. The "all snow" stations may have ratios in these figures which are in excess of 10.

Due to the local influence of bodies of water and the effects of altitude, interpolation between the plotted data on the charts must be done with care.

The ratio charts are intended to show in a rough way the frequency with which the snow-rain problem occurs at various places. In spite of the crudeness of the scheme, meteorologists familiar with the occurrence of borderline type precipitation have noted that the charts present a reasonable picture. Figure 10-24 illustrates one of these charts for the month of December.

You can readily see that by utilizing this type chart you would not only have an indication of the months and periods of greatest snowfall, but the percentage factor would give you a good indication of the likelihood of snow.

PHYSICAL NATURE OF THE PROBLEM

The type of precipitation that reaches the ground in a borderline situation is essentially

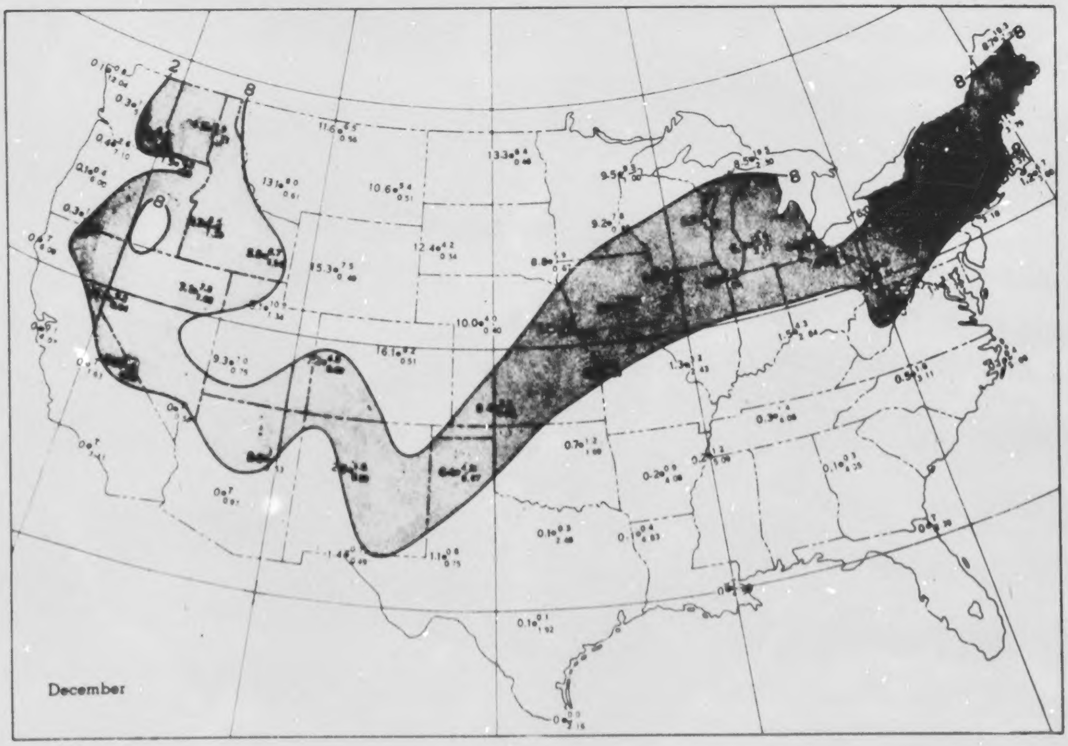

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Figure 10-24.–Ratio of normal monthly snowfall to normal total precipitation for the month of December. Large figures to the left of the station circle is the ratio, figure to the upper right is normal snowfall in inches for the period of record (1921-1950), and figure to the lower right is normal total precipitation in inches for the period of record. Regions where ratios are between 2 and 8 are shaded.

dependent upon whether there is a layer of above-freezing temperatures between the ground and the levels at which precipitation is forming, and whether this layer is sufficiently deep to melt all of the falling snow. Thus, the correct prediction of rain or snow at a given location depends largely upon the accuracy with which the vertical distribution of the temperature, especially the height of the freezing level, can be predicted. On the average, it is generally satisfactory to assume that the freezing level must be at least 1,200 feet above the surface to insure that most of the snow will melt before reaching the ground.

At the present state of forecasting capabilities, it is difficult to predict the entire sounding or even the height of the freezing level for periods of more than a few hours in the future. For periods of 12 to 36 hours, it is necessary to deal with much more general thermal parameters, such as temperatures at a few selected levels or mean temperatures of relatively thick layers. These will be discussed later in this chapter. The following is a review of the processes that influence local temperature changes, particularly those playing a critical role in the snow-rain problem.

Effects of Advection

In the lower troposphere (away from the immediate surface), horizontal advection is usually the dominant factor influencing the local temperature change. In most precipitation situations, particularly in the borderline situations, warm air advection and upward motion are both occurring so that warming is generally expected to accompany precipitation. These influences operate in the direction of turning the precipitation to rain, even if it starts as snow. However, this effect is frequently offset when there is weak warm advection or even cold advection in the cold air mass in the lower layers. The maintenance of subfreezing air in layers near the surface retards the effects of warming aloft; and even if the temperature of a significant layer aloft (about 1,200 feet or more thick) rises to values above freezing due to advection, there may be a prolonged period of sleet or freezing rain before the precipitation turns to rain.

In situations where precipitation is occurring in association with a cold upper low, upward motion is accompanied by little if any warm advection. In such borderline cases, precipitation may persist as snow, or will tend to turn to snow if it started as rain, due to cooling resulting from upward motion or advection.

Nonadiabatic Effects

The most important of the nonadiabatic effects is evaporational cooling which takes place as precipitation falls through unsaturated air between the clouds and the ground. This effect is especially pronounced when very dry air is present in low levels with wet-bulb temperatures at or below freezing. Then, even if the dry-bulb temperature is above freezing in a layer deeper than 1,200 feet in the lower levels, the precipitation may still fall as snow, since evaporation of the snow will lower the temperatures in the layer between cloud and ground until the below-freezing wet-bulb temperatures are approached.

The actual cooling which is observed during the period when evaporation is taking place is often on the order of 5° to 10°F within an hour or two. After the low level air is nearly saturated, evaporation practically ceases and advection brings a rise in the temperature in the low levels. However, reheating often comes too late to bring a quick change to rain since the temperatures may have dropped several degrees below freezing, much snow may have already fallen, and the lower levels may be kept cool through the transfer of any horizontally transported heat to the colder snow covered surface.

Melting of Snow

Melting of snow to rain in its descent through layers which are somewhat above freezing is another process which may cool the air in the lower troposphere. To obtain substantial temperature changes due to melting, it is necessary to have rather heavy amounts of precipitation falling and little warm air advection. As cooling proceeds far enough, the temperature of the entire lower sounding will reach freezing so that a heavy rainstorm can transform into a heavy snowstorm.

Cases of substantial lowering of the freezing level due to melting are relatively rare. This is probably due to the fact that the combination of heavy rain and little or no warm advection required for cooling due to melting is an infrequent occurrence.

Combined Effects

The combined effects of horizontal temperature advection, vertical motion, and cooling due to evaporation are well summarized by observations of the behavior of the bright band (melting layer) on radar. R. Wexler has observed that within the first 1 1/2 hours after the onset of precipitation, the bright band lowers by about 500 to 1,000 feet. This is attributable primarily to evaporational cooling and probably secondarily to melting. Since evaporational cooling terminates as saturation is reached, warm air advection (partially offset by upward motion) again becomes dominant and the bright band rises back to near its original level in about 3 hours after the onset of precipitation and may rise a few thousand feet more the end of 6 to 8 hours.

Other nonadiabatic effects, such as radiation and heat exchange with the surface, probably play a relatively smaller role in the snow-rain problem. However, it is likely that a difference in the state of the underlying surface (snow covered land vs open water) could determine whether the lower layers would be above or below freezing at a place in a particular precipitation situation. Occasionally along a seacoast in winter, heat from the open water keeps temperatures offshore above freezing in lower levels. Along the east coast of the United States, for example, immediate coastal areas may have rain, while a few miles inland, snow predominates. This is associated with low level onshore flow which is typical of many east coast cyclones. Actually this cannot be classified as a purely nonadiabatic effect relative to the land station at which rain rather than snow is occurring, since the warmer ocean air is being advected on shore and over the land involved.

GENERAL SYNOPTIC CONSIDERATIONS

The location of the snow-rain zone usually depends upon relatively small-scale synoptic considerations, such as the exact track of the surface disturbance, whether the wind at a coastal station is east or northeast, the position of the warm front, and the orientation of a ridge northeast of a low. The larger synoptic features do, however, determine the approximate position of the snow-rain zone. An awareness of this serves as an alert to the forecaster and is a useful tool in forecasts extending beyond 24 hours.

In the larger sense, the snow-rain zone is tied to the position of the polar front. The polar front location is in turn closely related to the position of the belt of strong winds in the middle and upper troposphere. When the westerlies are depressed southward, the storm track is similarly affected and the snow-rain zone may be as far south as the Southern United States. As the westerlies shift north of their normal position, the storm tracks develop across Canada. Concomitant with this northward shift, the United States has above normal temperatures, and the snow-rain problem may exist only along, or north of, the Canadian border.

With a flat, fast westerly flow aloft, the snow-rain zone will extend in a narrow westeast belt often well ahead of the surface perturbation and will undergo little latitudinal displacement as the perturbation moves across the country. Usually there is very little rain, and snow is found immediately north of the warm front. Most of the stations at which precipitation occurs will not undergo a change from one type of precipitation to another, since there is relatively little advection of warm or cold air with rapid zonal motion.

When the upper level wave is of large or increasing amplitude, it is difficult to generalize about the characteristics of the snow-rain problem without considering in detail the surface perturbation.

Up to now in this section, the snow-rain pattern has been discussed in association with an active low of the classical type. The rate of precipitation accumulation here is rapid and the transition period of freezing rain or sleet is short, usually on the order of a few hours or less. Another situation in which there is frequently a snow-rain problem is that of a quasi-stationary front in the Southern States with a broad west-southwest to southwest flow aloft and a weak surface low. The precipitation area in this case tends to become elongated in the direction of the upper level current. The precipitation rate may be slow, but occurs over a longer period. Often a broad area of sleet and freezing rain exists between belts of snow and rain, leading to a serious icing condition over an extensive region for a period of several hours or more. This pattern of precipitation changes either as an upper trough approaches from the west and initiates cyclogenesis on the front, or as the flow aloft veers and the precipitaiton dies out.

FORECASTING TECHNIQUES AND AIDS

Approaches to the snow vs rain forecasting problem have generally fallen into three broad categories. The first group depends on the use of the latest observed flow patterns and parameters derived therefrom to predict the prevalent type of precipitation for periods as much as 36 hours in advance. The second group consists of studies relating local parameters to the simultaneous occurrence of rain or snow at a particular station or area. In this approach, it is assumed that predicted values of the thermal parameter will be obtainable from circulation or other prognoses. Naturally, this approach tends to have its greatest accuracy for periods of 12 hours or less, since longer period temperature predictions for the boundary zone between rain and snow are very difficult to make with sufficient precision. A third approach used here involves the use of one of the many objective techniques available. A number of stations have developed objective techniques which are chiefly local in application. The method presented here is applicable to the eastern half of the United States. Thus, the general procedure in making the snow-rain forecast at present is to use a synoptic method (a purely subjective evaluation of the situation will have to suffice if an objective method has not been developed for the area in question) for periods up to 24 to 36 hours ahead, and then to consider expected behavior of thermal parameters over the area of interest to obtain more precision for periods of about 12 hours or less. The third step would be to employ any local objective technique. For a listing of some of the techniques available, consult the Air Weather Service Publication, Catalogue of Predictors Used in Local Objective Forecast Studies, AWSTR 105-19.

A number of methods based on synoptic flow patterns applicable to the United States are described in the U.S. Department of Commerce's publication, The Prediction of Snow vs Rain, Forecasting Guide No. 2. These methods are mostly local in application and are much too detailed to be presented in this training manual.

Prognostic charts from the National Meteorological Center and other sources should be utilized whenever and wherever available, not only to determine the occurrence and extent of precipitation, but for the prediction of the applicable thermal parameters as well.

Methods Employing Local Thermal Parameters

In this section we will discuss methods employing surface temperature, upper level temperatures, thicknesses, the height of the freezing level, and the forecast using combined parameters. All of these parameters are interdependent and should be considered simultaneously.

SURFACE TEMPERATURES.–The surface temperature by itself is not an effective criterion. Its use in the snow-rain problem has generally been used in combination with other thermal parameters. One study for the Northeastern United States found that, at 35°F, snow and rain occurred with equal frequency and by using 35°F as the critical value (predict snow at 35°F and below; rain above 35°F), 85 percent of the original cases could be classified. Another study based on data from stations in England suggested a critical temperature of 34.2°F and found that snow rarely occurs at temperatures higher than 39°F. However, it is obvious from these studies that even though surface temperature is of some general use in separating rain from snow, it is an inadequate discriminator in crucial cases. Thus, most investigators have looked to upper level temperatures as a further aid to the problem.

UPPER LEVEL TEMPERATURES.–Two studies of the Northeastern United States found that temperatures at the 850-mb level proved to be a good separating parameter and that including the surface temperature did not make any significant contribution. The separating temperatures at 850 mb were −2° to −4°C inclusive. Another study by J. T. Hilworth found that the area outlined by the 0°C isotherm at 550 mb and the 32°F isotherm on the surface chart, when superimposed upon the precipitation area, separated the types of precipitation in a high percentage of cases. The author's experience with this technique reveals that lower values should be used along coastal areas (in the −2° to −4°C category) and also behind deep cold lows. At mountain stations some higher level would have to be used.

A technique that utilizes temperatures from a standard isobaric surface is advantageous because these charts are usually available in the forecast office. There is, however, the difficulty that temperature inversions are occasionally located near the the 850- or 700-mb levels, so that the temperature of one level may not be indicative of any relatively deep layer of the atmosphere. This difficulty can be overcome by using thickness, which is a measure of the mean temperature of the layer.

THICKNESSES.–Thickness forecasting in relation to snow-rain studies has been used for a number of years. The National Meteorological Center has examined both 1,000-700 mb and 1,000-500 mb thickness limits for the eastern half of the United States. The critical values used are 9,200-9,400 feet (2,805-2,865 meters) for the 1,000-700 mb thickness and 17,600-17,800 feet (5,365-5,425 meters) for the 1,000-500 mb thickness. Also it has been noted that where the 1,000-500 mb thickness is 17,200 feet (5,243 meters) or less, the snow is more likely to be in the form of snow flurries than to be continuous snow.

Two studies in the United States, one for Fort Riley, Kansas, and the other for Mitchell Field, New York, revealed that the critical values or equal probability values for the 1,000-700 mb thicknesses were 9,350 feet (2,850 meters) and 9,250 feet (2,820 meters), respectively. Another study for Washington, D.C. suggested an upper limit of 9,350 feet for snow cases. The value of 9,300 feet for the eastern half of the United States appears to be a good average value of equal probability for the 1,000-700 mb thickness.

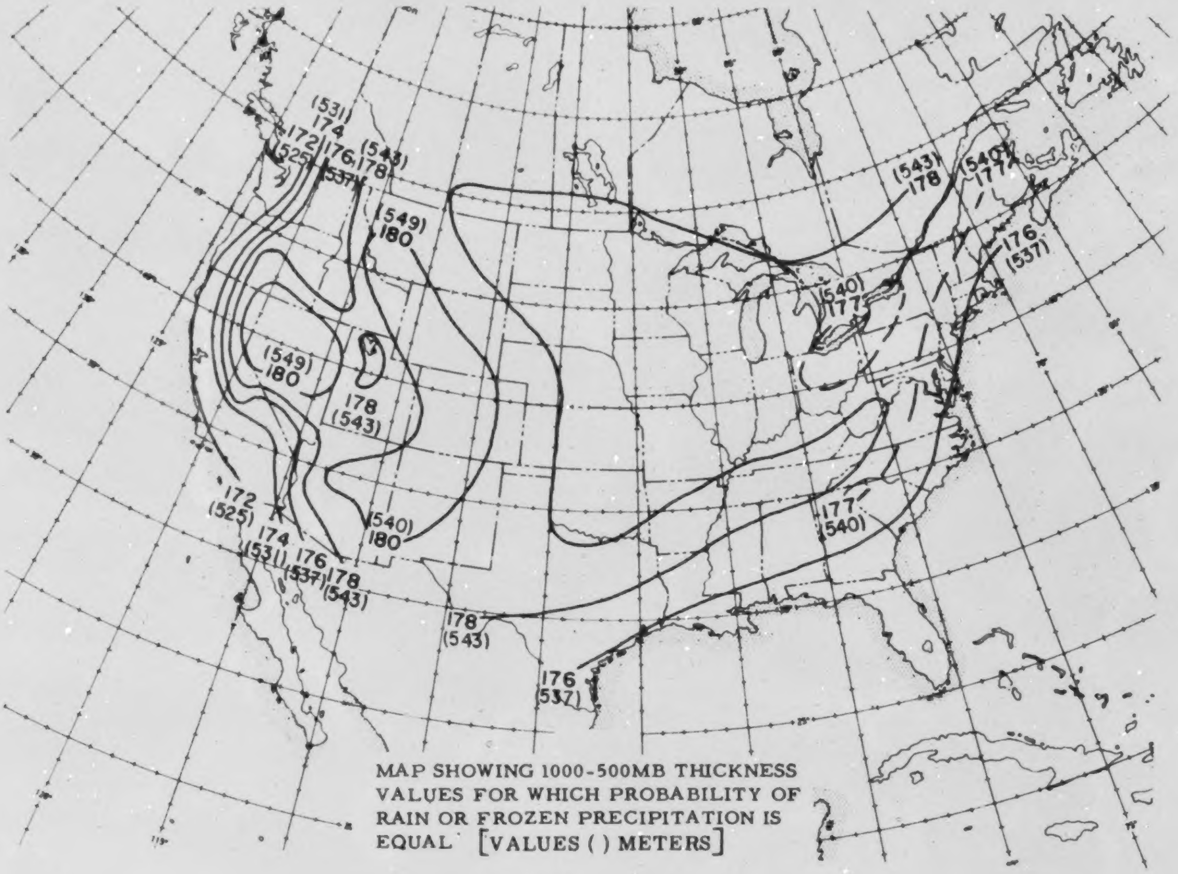
A more generalized study of 1,000-500 mb thickness as a predictor of the precipitation type in the United States was made by A. J. Wagner in 1957 on an Air Force GRD Contract. More complete details on this study may be found in The Prediction of Snow vs Rain, Forecasting Guide No. 2.

Wagner's study was taken from data from 40 stations over the United States for the colder months of a 2-year period. Cases were limited to surface temperatures between 10°F and 50°F. The type of precipitation in each case was considered as belonging in one of two categories, as follows: Frozen, which includes snow, sleet, granular snow, and snow crystals; and unfrozen, which includes rain, rain and snow mixed, drizzle, and freezing rain and drizzle.

Equal probability or critical thickness values were obtained from the data at each station. From this study it is clear that the critical thickness increases with increasing altitude. This altitude relationship is attributable to the fact that a sizable part of the thickness layer is nonexistent for high-altitude stations and obviously does not participate in the melting process. In order to compensate for this, the equal probability thickness must increase with station altitude. For higher altitude station's thicknesses between 850-500 mb or 700-500 mb, as appropriate, should prove to be better related to precipitation type.

The Wagner equal probability chart is reproduced in figure 10-25.

Wagner's study also indicates that the type of precipitation can be specified with a certainty of 75 percent at plus or minus 100 feet (30 meters) from the equal probability value, increasing to 90 percent certainty at plus or minus 300 (90 meters) feet from this value. Stability is the parameter that accounts for the variability of precipitation for a given thicknes at a given point. This fact is taken into account in the following manner. For example, if the forecast precipitation is due to a warm front which is much stronger (more stable) than usual, the line separating rain from frozen precipitation is shifted toward higher thickness values. Over the Great Lakes, where snow occurs in unstable or stable conditions (frontal), the equal probability thickness is lower than that shown in figure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MAP SHOWING 1000-500MB THICKNESS VALUES FOR WHICH PROBABILITY OF RAIN OR FROZEN PRECIPITATION IS EQUAL [VALUES ( ) METERS]

AG.599

Figure 10-25.—Map showing 1,000- 500-mb thickness values for which probability of rain or frozen precipitation is equal (after Wagner).

10-25 for snow showers, and higher than that shown in figure 10-25 for warm frontal snow.

In using this technique over the United States, the current transmission of thickness progs can be utilized. Since shorter period forecasts of thicknesses are more accurate, it is likely that the results of these studies of thickness related to precipitation type will be of the greatest assistance in snow-rain prediction for periods of 12 hours or less.

HEIGHT OF THE FREEZING LEVEL.—The height of the freezing level above the surface is one of the most critical thermal parameters in determining whether snow can reach the ground. It was pointed out earlier in this section of the chapter that theoretical and observational evidence indicates that a freezing level averaging 1,200 feet or more above the ground is usually needed to insure that most of the snow will melt before reaching the ground. This figure of 1,200 feet can thus be considered as a critical or equal probability value of the freezing level. In practice, since prediction of the freezing level is rather difficult, even for a short period of time ahead, it has rather limited value as a predictor of rain vs snow.

COMBINED THERMAL PARAMETERS.—From the foregoing discussion, it is concluded that no one method, when used alone, is a good discriminator in the snow-rain forecasting problem. Therefore, it is suggested that the combined use of the surface temperature, height of the freezing level, 850-mb temperature, and the 1,000-700 and/or 1,000-500 mb thicknesses be made to arrive at the forecast. There is generally a high correlation between the 850-mb temperature and the 1,000-700 mb thickness and between the 700-mb temperature and the 1,000-500 mb thickness. Certainly an accurate temperature forecast for these two levels would yield an approximate thickness value for discriminating purposes.

Hilworth Method

This technique and the following technique on forecasting the area of maximum snowfall were taken from Further Studies in the Development of Short Range Weather Prediction Techniques, GRD Contract No. AF19(604)-2073 Scientific Report No. 1, by J.J. George and Associates.

The determining factor in the type of precipitation in this study was found to be the distribution of temperature and moisture between the surface and the 700-mb level at the time of beginning of precipitation. However, prediction of the sounding of this strata with any degree of accuracy was found to be quite involved and impractical. Therefore, the median level of 850 mb was studied in conjunction with the precipitation area and the 32°F isotherm sketched on the surface synoptic chart. This method presents an objective, yet practical method, by which the forecaster can make a decision on whether the precipitation in winter will be rain, snow, freezing rain, sleet, or some combination of these.

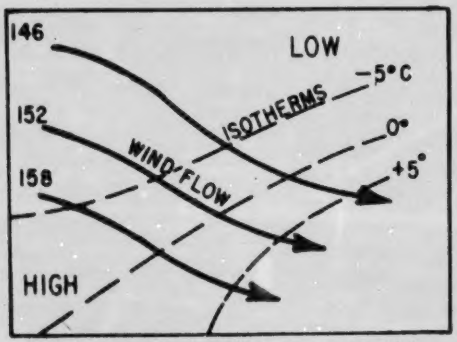
The following objective techniques can be applied to the land area south of 50° north latitude, and east of a line drawn through Williston; North Dakota; Rapid City, South Dakota; Goodland, Kansas; and Amarillo, Texas.

It was found that the area outlined by the 0°C isotherm at 850 mb and the 32°F isotherm on the surface chart, when superimposed upon the precipitation area, generally separates the types of precipitation; that is, most of the pure rain was found on the warm side of the 32°F isotherm, and most of the pure snow on the cold side of the 0°C isotherm with intermediate types falling generally within the enclosed area between these two isotherms. It was further realized that in a large majority of situations, evaporation and condensation was a sizable factor, both at 850 mb and at surface levels in its effect upon temperature. With this in mind, the wet-bulb temperature was selected for investigation because of its conservative properties with respect to evaportaion and condensation, and also because of its ease of computation directly from the temperature and dewpoint. Graphs are shown in the following section, along with rules for movement of the 850-mb temperature and 0°C wet-bulb isotherm. The surface chart is used for computations of the 1,000-mb level since the surface chart approximates the 1,000-mb level for most stations during a snow situation and therefore little error is introduced. IT MUST BE REMEMBERED THAT ALL PREDICTIONS ARE BASED ON FORECAST VALUES.

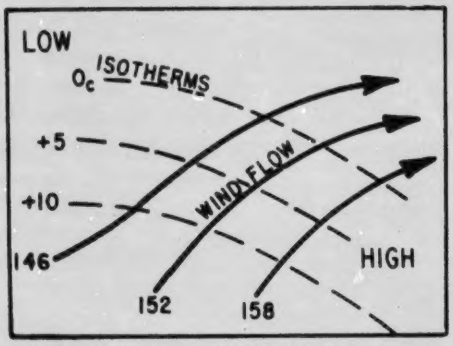
MOVEMENT OF THE 850-MB 0°C ISOTHERM. No completely objective method of forecasting the 850-mb isotherms is available. A reasonably good approximation can be made subjectively by the use of the following rules and by a combination of extrapolation and advection, tempered with synoptic developments. (See fig. 10-26(A) and (B) for typical warm and cold air advection pattern at 850 mb.)

The following rules for the movement of the 24-hour 850-mb temperature change areas have been devised:

1. Maximum cooling takes place between the 850-mb contour trough and the 850-mb isotherm ridge east of the trough.
2. Maximum warming takes place between the 850-mb contour ridge and the 850-mb isotherm trough east of the contour ridge.
3. Changes are slight with an ill-defined isotherm and/or contour pattern.
4. Usually, little change occurs when isotherms and contours are in phase at the 850-mb level.
5. The temperature falls at 850 mb tend to replace height falls at 700 mb in an average of 24 hours. Conversely, temperature rises replace height rises.



(A) COLD AIR ADVECTION (850mb)



(B) WARM AIR ADVECTION (850mb)

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Figure 10-26.—Typical cold and warm air advection patterns at 850 mb. (A) Cold; (B) Warm.

6. With filling troughs or northeastward moving lows, despite northwest flow behind the trough, 850-mb isotherms are seldom displaced southward, but follow the trough toward the east or northeast.

7. Always predict temperature falls immediately following a trough passage.

8. Do not forecast temperature rises of more than 1 or 2 degrees in areas of light or sparse precipitation in the fore-trough. If the area of precipitation is widespread and moderate or heavy, forecast no tempetature rise.

9. Upslope effect in the United States starts approximately at the 100th meridian for southeast winds. Such a pattern will result in slight 24-hour cooling where warming might otherwise be indicated.

10. With eastward moving systems under normal winter conditions (troughs at 700 mb moving east about 11° per day), a distance of 400 nautical miles to the west is a good point to locate the temperature to be expected at the forecasting point 24 hours hence. A good 850-mb temperature advection speed seems to be about 75 percent of the 700-mb trough displacement.

The following is a step-by-step procedure for moving the 850-mb 0°C isotherm:

1. Extrapolate for 12 and 24 hours the thermal ridge and trough points. If poorly defined, this step may be omitted. The amplitude of the thermal wave may be increased or decreased subjectively if, during the past 12 hours, there has been a corresponding increase or decrease in the height of the contours at 500 mb.

2. The thermal wave patterns will maintain the approximate relative position with the 850-mb height troughs and ridges. Therefore, the 12- and 24-hour prognostic position of the contour trough and ridges should be made and the extrapolated positions of the thermal points checked against this contour prog. Adjustments of these points should be made.

3. Select points on the 0°C isotherm that lie between the thermal ridge and trough as follows: one or two in the apparent warm advection area, and one or two in the apparent cold advection area. Apply the following rules to these selected points.

a. Warm Advection Area. If the point lies in a near saturated or precipitation area, it will remain practically stationary with respect to the contour trough. If the point lies in a nonsaturated area but one that is expected to become saturated or actually to lie in precipitation area, then it will remain stationary or move upwind slightly to approximately the prognostic position of the 0°C wet-bulb. If the point does not fall in the above two categories, it will advect with about 50 percent of the wind component normal to the isotherm. Note in all three cases above the movement is related to the contour pattern.

b. Cold Advection Area. Advect the point with approximately 75-80 percent of the wind component normal to it.

4. In the case of a closed low at 850 mb moving slowly, the 0°C isotherm will move eastward with respect to the closed low as cold air is advected all the way around the low.

MOVEMENT OF THE 850-MB 0°C WET-BULB ISOTHERM.–The wet-bulb temperature can be forecast by the above procedure and rules in a general sense, remembering that it is dependent upon dewpoint as well as the temperature. The dewpoint will advect with the winds at nearly the full velocity, whereas the temperature under nonsaturated conditions moves slower. As saturation is reached, the necessity of treatment separate from the 0°C isotherm disappears. The following observations with respect to the 0 ° C wet-bulb isotherm may help:

1. The 0°C wet-bulb isotherm does not move far offshore in the Gulf and the Atlantic, because of convection in the cold air over warm water.
2. If the 0°C wet-bulb isotherm lies in a ribbon of closely packed isotherms, movement is slow.
3. Extrapolation works well on troughs and ridges.

METHOD OF APPLICATION.–After the forecast of the surface and 850-mb level temperature and dewpoint values are made, you are ready to convert these values to their respective wet-bulb temperatures. The following procedure is recommended:

1. Using figure 10-27(A) and (B), compute the wet-bulb temperatures for the 850- and 1,000-mb levels, respectively. (The surface chart is used for the 1,000-mb level.) Admittedly, the wet-bulb temperatures at just these two levels does not give a complete picture of the actual distribution of moisture and temperature, and error is introduced when values are changing rapidly, but these are values the forecaster can work with and predict with reasonable accuracy.
2. Enter these values on the worksheet, the bottom of figure 10-28. Then use the top of this figure with your predicted values to obtain the forecast. A necessary assumption for use of this graph is that the wet-bulb temperatures at these two levels can be predicted with reasonable accuracy. Known factors affecting the wet-bulb temperature at any particular station should be carefully considered before entering the graph. Some of the known factors are elevation, proximity to a warm body of water, known layers of warm air above or below 850 mb, etc. Area "A" on the graph calls for a rain forecast, area "B" for a freezing rain forecast, and area "C" for a snow forecast. Area "D" is not so clear cut, being an overlap portion of the graph; however, wet snow or rain and snow mixed predominate in this area. Sleet occurring by itself for more than 1 or 2 hours was found to be rare and should be forecast with caution. (See figure 10-28 for sample values and snow forecast.)

Forecasting the Area of Maximum Snow

The basic intent of this section of the chapter is the prediction of the area of maximum snowfall. This classification does not include snowstorms which can be classified as air mass, or purely local type storms attributable to features of the terrain. In this category are those severe snowstorms which occur to the lee of the Great Lakes under more or less steady west-to-north flow of polar air. On the other hand, snow cases occurring in connection with active lows or associated with areas of vertical sheer are included.

SYNOPTIC TYPES.–Four distinct types of synoptic patterns with a maximum snow area associated with it are discussed below.

BLIZZARD TYPE.–The synoptic situation features an occluding low pressure center. In the majority of cases the "wrapped around" high pressure and ridges are present. The track of the low is north of 40° and its speed, which initially may be average or about 25 knots, decreases into the slow category during the occluding process. In practically all cases a cold closed low at 500 mb is present and captures the surface low in 24-36 hours.

The area of maximum snowfall lies to the left of the track. At any particular synopitc position, the area is located from due north to west of the low center. Within this maximum area the rate of snowfall would be classified moderate (1/2 to 1 inch per hour) in most cases. However, when this type occurs on the east coast with its large

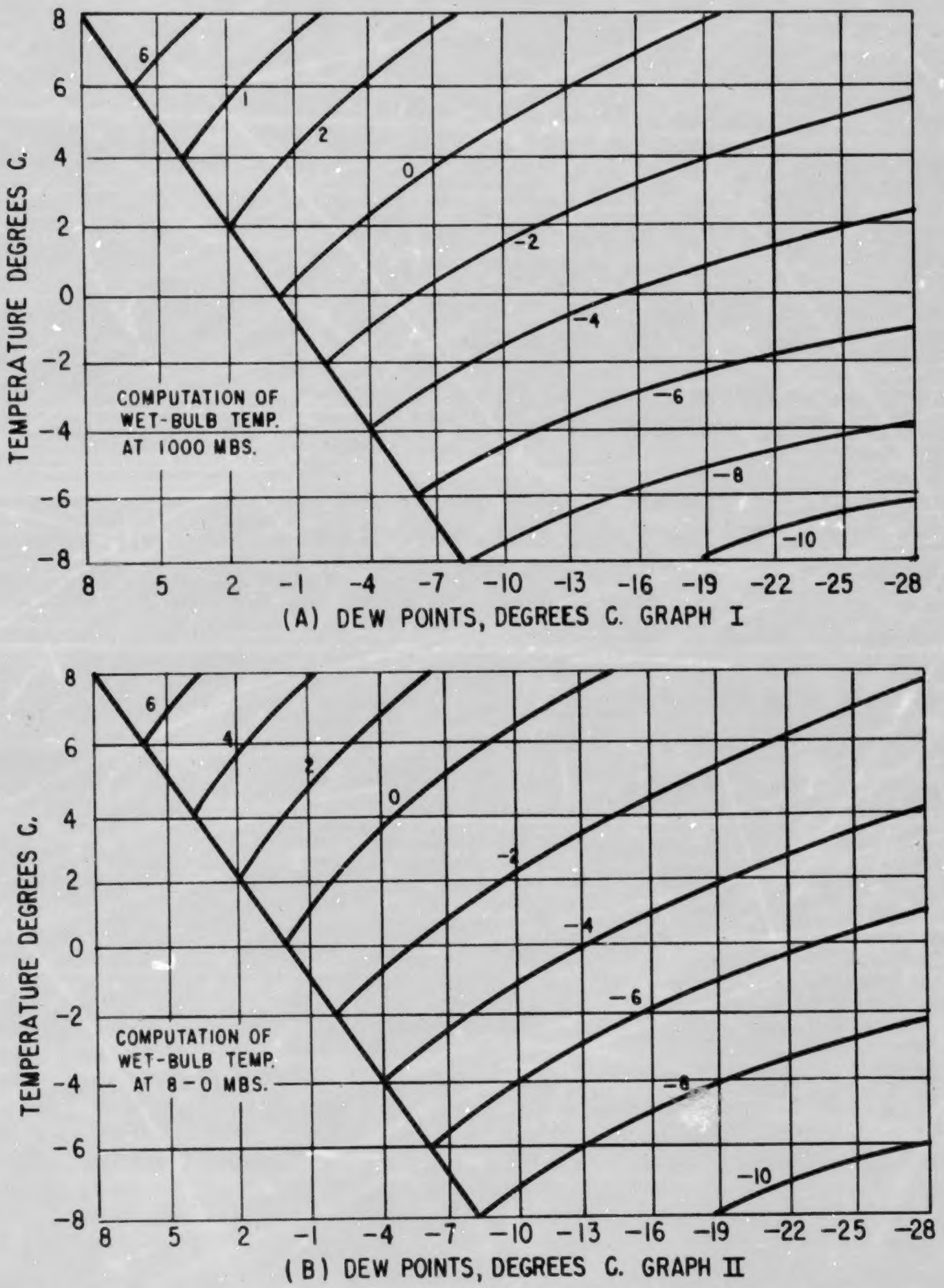

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Figure 10-27.—Graphs for computing wet-bulb temperatures. (A) Computation of 1,000-mb wet-bulb temperature; (B) computation of 850-mb wet-bulb temperature.

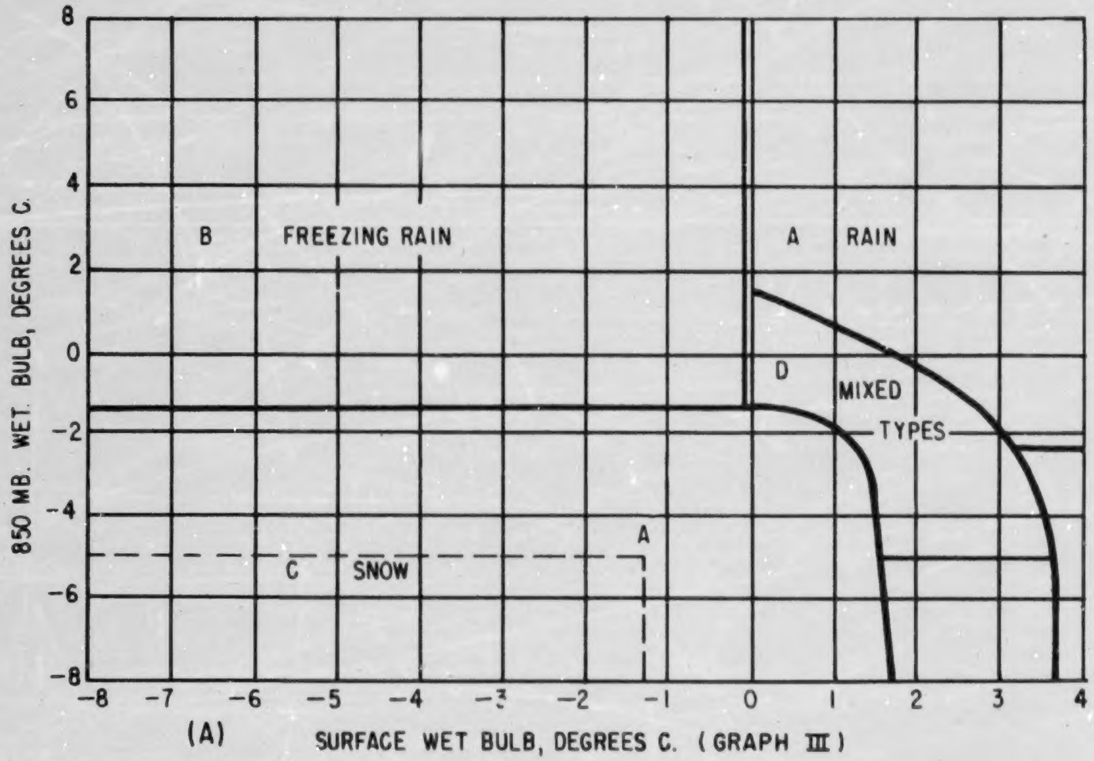

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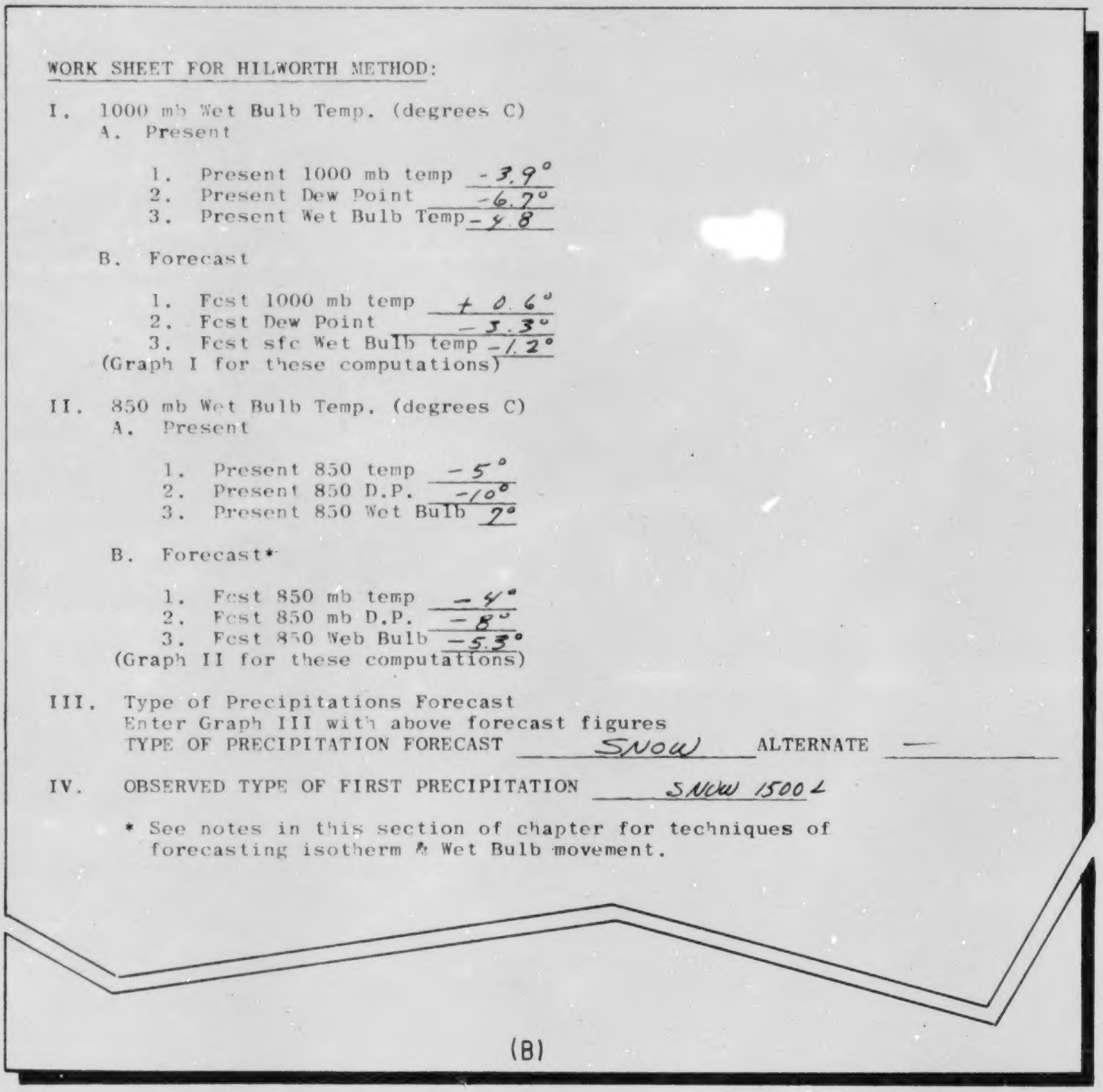
Figure 10-28.–Graph (A) and worksheet (B) for delineating the type of precipitation using the Hilworth method. (Point A is intersection of forecast values–snow is forecast.)

temperature contrast and high moisture availability, the heaviest known snowfalls occur. The western edge of the maximum area is limited by the 700-mb trough or low center, and the end of all snow occurs with the passage of the 500-mb trough or low center. Therefore, it follows that the maximum snow area diminishes and contracts during the capturing process. Often a new snow area forms to the east in connection with cyclogenesis or center jump; otherwise, the snow area, now confined to the vicinity of the surface low and considerably diminished in intensity, moves off to the north or northeast.

MAJOR STORM AND NONOCCLUDING LOWS.–The synoptic situation consists of a wave type low of the nonoccluding type. In practically all cases, the "wrapped around" high pressure and ridges are present. The track of the low or wave is east of 40° and its speed is at least the average of 25 knots, often falling into the fast moving category. The upper air picture is one of fast moving troughs, generally open, but on occasion could have a minor closed center for one or two maps in the bottom of the trough.

The area of maximum snowfall lies in the cold air to the left of the track of the low and usually describes a narrow belt oriented east-west or northeast-southwest about 100-200 miles wide. At any particular synoptic position the area is located parallel to the warm front and north of the low center. Within the maximum area, the rate of snowfall is variable from one case to another, depending upon available moisture, amount of vertical shear, etc. However, it is not uncommon for heavy snow (1 inch per hour or



WORK SHEET FOR HILWORTH METHOD:

I. 1000 mb Wet Bulb Temp. (degrees C)
  A. Present
    1. Present 1000 mb temp -3.9°
    2. Present Dew Point -6.7°
    3. Present Wet Bulb Temp -4.8

  B. Forecast
    1. Fcst 1000 mb temp +0.6°
    2. Fcst Dew Point -3.3°
    3. Fcst sfc Wet Bulb temp -1.2°
  (Graph I for these computations)

II. 850 mb Wet Bulb Temp. (degrees C)
  A. Present
    1. Present 850 temp -5°
    2. Present 850 D.P. -10°
    3. Present 850 Wet Bulb 7°

  B. Forecast*
    1. Fcst 850 mb temp -4°
    2. Fcst 850 mb D.P. -8°
    3. Fcst 850 Web Bulb -5.3°
  (Graph II for these computations)

III. Type of Precipitations Forecast
  Enter Graph III with above forecast figures
  TYPE OF PRECIPITATION FORECAST SNOW ALTERNATE —

IV. OBSERVED TYPE OF FIRST PRECIPITATION SNOW 1500L

* See notes in this section of chapter for techniques of forecasting isotherm & Wet Bulb movement.

(B)

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Figure 10-28.—Graph (A) and worksheet (B) for delineating the type of precipitation using the Hilworth method. (Point A is intersection of forecast values—snow is forecast.)—Continued.

greater) to occur. It must be remembered that even though heavy snow does occur, the duration is short by the very nature of the storm. This means that a station would lie in the maximum snow area only 4 to 8 hours; whereas in the case of the blizzard type, it usually remains in the area in excess of 10 hours.

WARM ADVECTION TYPE.—This type, which occurred only a few times in the initial study, was separated from the other types because of the absence of an active low in the vicinity of the maximum snow area. A blocking high-pressure ridge or wedge is present ahead of a sharp warm front. The overrunning warm air is a steady current from the south to southwest. The area of maximum snowfall is a narrow band parallel to the warm front and moves north or northeast. Only under the most ideal conditions does this area become serious: nearly stationary front, ample supply of moisture, and persistent flow aloft. A rate of fall of moderate to heavy for a 6- to 12-hour duration may occur. The usual history is a transition to freezing rain, then rain.

POST-COLD FRONTAL TYPE.–The synoptic situation consists of a sharp cold front oriented nearly north-south in a deep trough. A minor wave may form on the front and travel rapidly north or northeast along it. Strong cold advection from the surface to 850 mb is present west of the front. The troughs at 700 mb and 500 mb are sharp and displaced to west of the surface trough 200-300 miles. Ample moisture is available at 850 mb and 700 mb. This type of heavy snow area occurs once or twice a season.

The area of maximum snowfall is located between the 850-mb and 700-mb troughs where moisture at both levels is available. The rate of fall is moderate, although for a brief period of an hour or less it may be heavy. The duration is short, of the order 2 to 4 hours at any one station. The area as a whole generates and dies out in a 12- to 18-hour period. The normal history is one of a general area of light snow within the first 200 miles of a strong push of cold air. After the cold air moves far enough south and the cold front becomes oriented more N-S and begins moving eastward steadily, the troughs aloft and moisture distribution reach an ideal state and a maximum snow area appears. After 12-18 hours the advection of dry air at 700 mb decreases the rate or fall in the area, and soon thereafter the areas as a whole dies out.

LOCATING AREA OF MAXIMUM SNOWFALL.–TEMPERATURE. The 0°C (-3°C east coast) isotherm at 850 mb is used as the basic defining line for the snow area. This isotherm should be carefully analyzed, using all data at 850 mb. It should then be checked against the surface map, keeping in mind the following points:

1. In areas of precipitation, stations reporting snow should lie on the cold side of the 0°C (-3°C east coast) isotherm; stations reporting mixed types of precipitation (e.g., rain and snow, sleet and snow), the 0°C isotherm will lie very close to or through the station.

2. In areas of no precipitation, the 0°C (-3°C east coast) isotherm will roughly parallel the 32°F isotherm at the surface. In cloudy areas the separation will be small, and in clear areas the separation will be larger.

At the 850-mb level the 0°C wet-bulb temperature should be sketched in, particularly in the area where precipitation may be anticipated within the next 12-24 hours. This line will serve as the first approximation of the future position of the 0°C isotherm.

MOISTURE.–At the 850-mb level the -5°C dewpoint line, and at 700-mb level, the -10°C dewpoint line are used as the basic defining lines. The area at 850 mb that lies within the overlap of the 0°C isotherm and the -5°C dewpoint line is the first approximation of the maximum snowfall area. All stations within this area have temperatures less than 0°C and spreads of 5°C or less. This area is then further refined by superimposing the sketched -10°C dewpoint line at 700 mb upon the area. Now the final area is defined by the 0°C isotherm and the overlapped minimum dewpoint lines from both levels. This final area becomes the area where moderate or heavy snow will be reported, depending upon the particular synoptic situation. (See fig. 10-29.)

MOVEMENT.–As might be supposed, the area of maximum snowfall is in motion and must be forecast. The first basic rule for moving the area is that it maintains the same relative position to the other synoptic features of the 850-mb level and surface chart. However, in order to forecast the expansion or contraction of the area, it is necessary to forecast the lines that define it. The 0°C isotherm should be forecast according to the rules set forth in the section treating this particular phase. the moisture lines may be advected with the winds as set forth previously in this chapter. The 0°C isotherm should also be moved with rules stated previously in this chapter.

It was found that the area of maximum snowfall could be forecast for 12 hours with

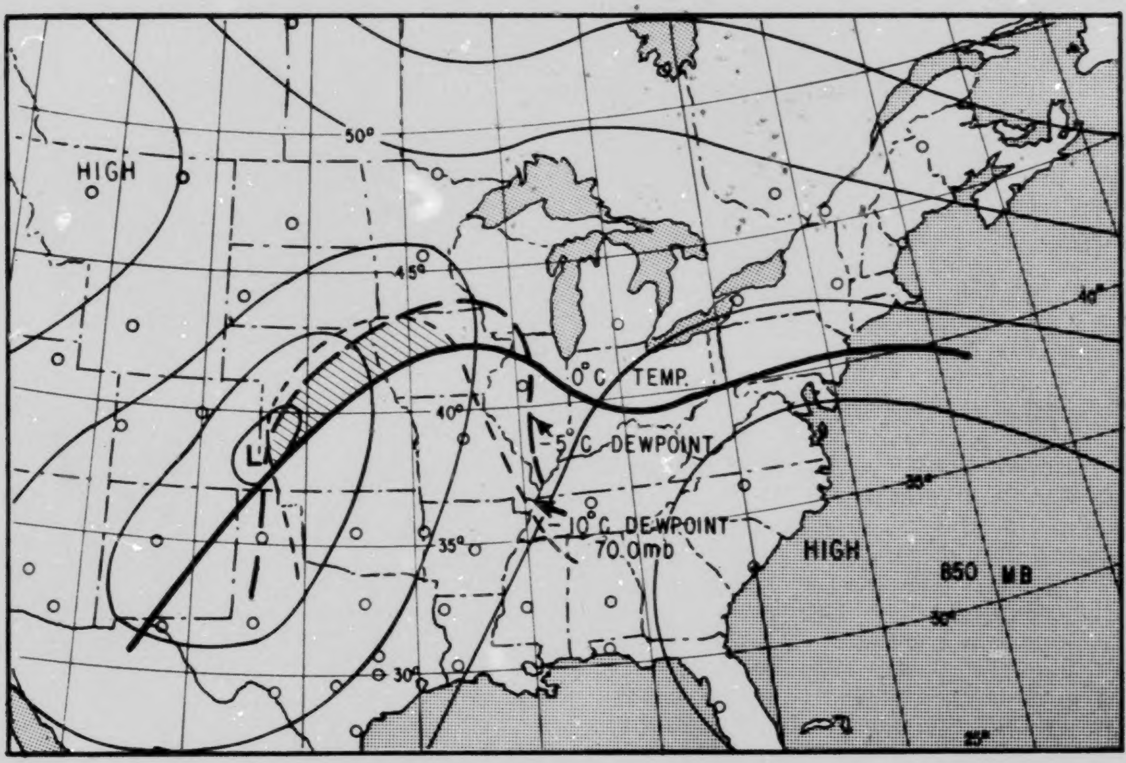

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Figure 10-29.—Illustration of the location of the maximum snow area. The low center moved to Iowa in 24 hours, and the maximum snow area spread northeast along the area 50-75 miles either side of a line through Minneapolis to Houghton, Michigan.

considerable accuracy and for 24 hours with fair accuracy, provided a reasonable amount of care was exercised according to rules and subjective ideas mentioned above.

QUANTITATIVE PREDICTION.–The National Weather Analysis Center currently makes quantitative precipitation predictions and transmits them on the National Weather Facsimile Network. Details on this procedure may be found in the U.S. Department of Commerce publication, Synoptic Meteorology as Practiced by the National Meteorological Center, The NMC Manual, Part II, NW 50-1P-548, latest revision. These predictions are based on vertical motion, precipitable water, humidity, and stability.

For local forecasting you should not only consult these forecasts but make your own evaluation as well. Factors to be considered should be the climatology of snow for your particular area, the type and intensity of the storm producing the precipitation, its speed and duration over your station, and the relation of your station to the area of maximum snowfall. One rule of thumb which has been in use for several years involves the use of the amount of precipitable water from values found on the chart transmitted over the National Weather Facsimile Network. This rule uses a ratio of 1 inch of precipitation equals 10 inches of snow. Therefore, using this rule of thumb, the maximum amount of snowfall which could occur would be in direct ratio to the amount of precipitable water overlying the station. However, you must consider moisture advection and changing values. For example, if your station showed 1/2 inch of precipitable water at present with no significant change during the forecast period, and the precipitation forecast was snow, 5 inches of snow would be the maximum you would expect. This rule should be used with caution and in conjunction with all other available information.

APPLICATION TO LOCAL AREA

Figure 10-30 presents a composite worksheet which is a suggested format for correlating the thermal parameters and techniques presented in the foregoing section. You must remember that this worksheet is based on the assumption that the synoptic type, climatology, and other indications reveal that precipitation will occur and that there is a likelihood of the precipitation being in the form of snow or some other frozen type. Also, all of the values used in these techniques are FORECAST VALUES. Local predictions, facsimile prognostic charts, and other available data should be utilized to determine the forecast values and parameters at the time of the ONSET of precipitation at your station or locality. The example given is a fictitious station and merely presents an illustration of the proper step-by-step procedure which can be used to take all of the parameters into consideration. A worksheet of this type could be prepared for your local station using the optimum values.

APPLICATION TO A LARGE AREA

Frequently, flight operations reach into an area outside the local sphere of influence. In these cases, and for other reasons, it may be desirable to have a rough delineation of the snow and rain areas on the synoptic map. For a rough approximation of the snow-rain dividing line over a relatively large area, the following techniques are recommended:

1. On the thickness chart and on the thickness prog, draw both the 17,800- and 17,600-foot (5,425 and 5,365 meters) thickness lines. If an overlay acetate chart is available, this can be placed in register with the synoptic chart to determine the snow-rain dividing line. In the majority of cases (over the eastern part of the United States) these lines will divide the snow-rain areas as follows: rain will be found on the higher side of these lines, and snow or some frozen form of precipitation on the lower side. Be sure to keep in mind the rules set forth previously in this chapter. (Also use the Wagner equal probability chart.)

2. The George method can also be utilized as a further check. This method was outlined previously. Use the 32°F line on the surface and the 0°C line on the 850 mb chart to divide the areas of solid and liquid precipitation. In some cases the use of the −3°C line at 850 mb is more desirable.

3. For determination of the areas of maximum snowfall, use the methods previously explained, keeping in mind the synoptic types.

TEMPERATURE

Temperature ranks among the most important forecast elements. Temperatures are not only important for some operational procedures but also are of keen interest to all of us in everyday life.

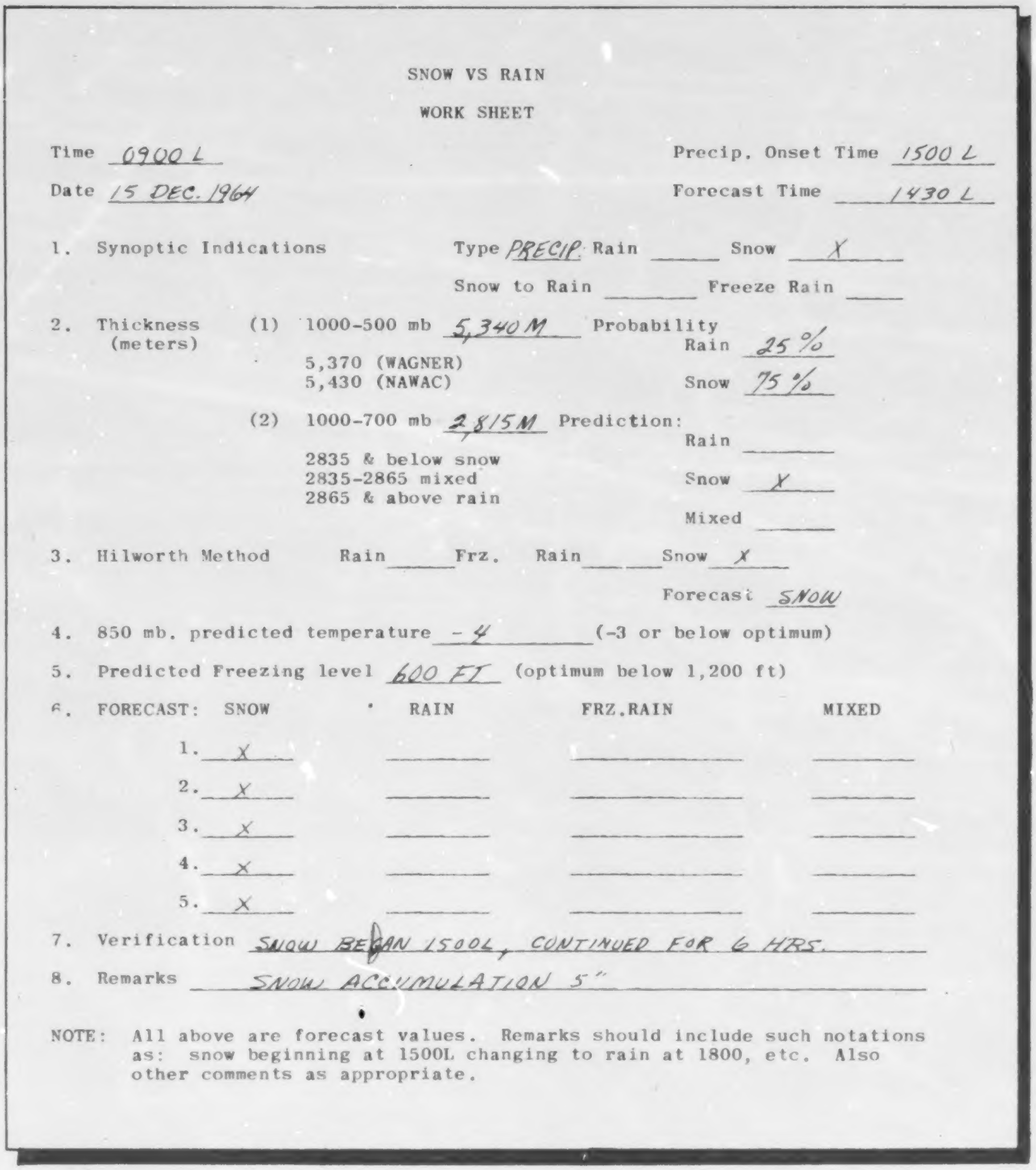
FACTORS AFFECTING TEMPERATURES

In forecasting temperatures, many factors are involved. These factors include air mass characteristics; frontal positions, characteristics, and movement; amount and type of cloudiness; season; nature and position of pressure systems; and local conditions.

Temperature, being subject to marked changes from day to night, is not considered a conservative property of an air mass. Too, it does not always have a uniform lapse rate from the surface up through the atmosphere. This means that the surface air temperature will not be representative because of the existence of an inversion, a condition particularly prevalent at night. Usually the noonday surface air temperature is fairly representative.

Let us look at the factors which cause temperature variations: Insolation and terrestrial radiation, lapse rate, advection, vertical heat transport, and evaporation and condensation.

In forecasting temperature, consider insolation and terrestrial radiation as two very important factors. Low latitudes, for instance, receive more



SNOW VS RAIN

WORK SHEET

Time 0900 L

Precip. Onset Time 1500 L

Date 15 DEC. 1964

Forecast Time 1430 L

1. Synoptic Indications Type PRECIP. Rain ____ Snow X

Snow to Rain ____ Freeze Rain ____

2. Thickness (meters) (1) 1000-500 mb 5,340 M Probability

5,370 (WAGNER)
5,430 (NAWAC)

Rain 25%

Snow 75%

(2) 1000-700 mb 2,815M Prediction:

2835 & below snow
2835-2865 mixed
2865 & above rain

Rain ____

Snow X

Mixed ____

3. Hilworth Method Rain ____ Frz. Rain ____ Snow X

Forecast SNOW

4. 850 mb. predicted temperature -4 (-3 or below optimum)

5. Predicted Freezing level 600 FT (optimum below 1,200 ft)

6. FORECAST:

	SNOW	RAIN	FRZ.RAIN	MIXED
1.	X			
2.	X			
3.	X			
4.	X			
5.	X			

7. Verification SNOW BEGAN 1500L, CONTINUED FOR 6 HRS.

8. Remarks SNOW ACCUMULATION 5"

NOTE: All above are forecast values. Remarks should include such notations as: snow beginning at 1500L changing to rain at 1800, etc. Also other comments as appropriate.

AG.605

Figure 10-30.—Sample composite worksheet for the snow versus rain forecasting problem.

heat during the day than stations at high latitudes. More daytime heat can be expected in the summer than in the winter, since in the summer the sun's rays are more direct and reach the earth for a longer period. Normally, there is a net gain of heat during the day and a net loss at night. Consequently, the maximum temperature is usually reached during the day; the minimum, at night. Cloudiness will affect insolation and terrestrial radiation. Temperature forecasts must be made only after the amount of cloudiness expected is determined. Clouds reduce insolation and terrestrial radiation, causing daytime temperature readings to be relatively lower than normally expected and nighttime temperatures to be relatively higher. The stability of the lapse rate has a marked effect on insolation and terrestrial radiation. With a stable lapse rate there is less vertical extent to heat; surface heating therefore takes place more rapidly. With an unstable lapse rate, the opposite is true. If there is an inversion, there is less cooling, since the surface temperature is lower than that of the inversion layer; that is, at some point the energy radiated by the surface is balanced by that radiated by the inversion layer.

One of the biggest factors affecting temperature is the advection of air. Advection is particularly marked in its effect on temperature with frontal passage. If a frontal passage is expected during the forecast period, the temperature must be considered. The temperature gradient with an air mass may be as important as frontal passage in forecasting. Advection within an air mass may also be important. This is particularly true of sea and land breezes and mountain breezes. They affect the maximum and minimum temperatures and their time of occurrence.

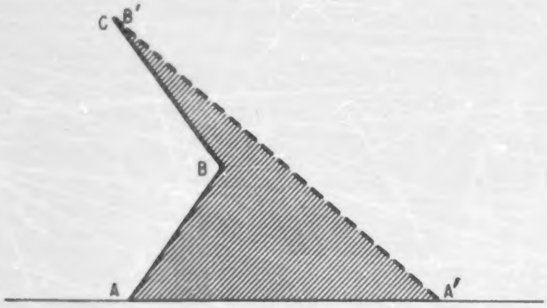
Vertical heat transport is a temperature factor. It is considerably affected by the speed of the wind. With strong wind there is less heating and cooling than with light wind or a calm since the heat energy gained or lost is distributed through a deeper layer when the turbulence is greater.

Evaporation and condensation affect the temperature of an air mass. When cool rain falls through a warmer air mass, evaporation takes place, taking heat from the air. This often occurs to affect the maximum on a summer day on which afternoon thundershowers occur. The temperature may be affected at the surface by condensation to a small extent during fog formation, raising the temperature a degree or so because of giving off the latent heat of condensation to the air. Refer to an earlier section of this chapter for a more detailed discussion on evaporation and condensation effects on temperature.

FORECASTING MAXIMUM TEMPERATURE BY USE OF SKEW T LOG P DIAGRAM

The Skew T Log P diagram may be used to get a good approximation of the maximum temperature by plotting the most recent meteorological sounding on the chart. If there is a surface inversion on the sounding, the approximate maximum temperature may be found on the diagram by following down the dry adiabat from the top of the inversion to the surface. This approximates the maximum temperature, since any heating beyond this point will be spread through a large vertical section. The more the lapse rate above the inversion approximates the dry adiabat, the nearer this temperature will approximate the maximum temperature.

Another method of forecasting the maximum temperature is by the use of the EQUAL AREA method. On figure 10-31 if ABC represents the sounding at the time of the minimum temperature on the first day and A'B'C represents the sounding at the time of the maximum temperature, the hatched area represents the energy added by heating during the day. If the air mass, wind, and cloudiness conditions remain the same from one day to the next, the same amount of insolation should be received. Consequently, if the sounding at the time of the minimum temperature on the second day is used and an energy area equal to that of the previous day is added to it, the maximum temperature for the day should be indicated. Normally, the actual meteorological sounding is not made at the time of the maximum or minimum temperature; therefore, to get the best results, the actual sounding should be modified to fit the maximum or minimum temperature. However, use of the actual morning sounding gives a fair



AG.606

Figure 10-31.—Energy area for forecasting maximum temperature.

approximation. One of the disadvantages of this method is that it gives a forecast value for only one day.

FORECASTING MINIMUM TEMPERATURE BY USE OF SKEW T DIAGRAM

The equal area method of forecasting may also be employed in arriving at a forecast of the minimum temperature. Instead of adding the energy area, as was the case in forecasting the maximum temperature, the energy area is subtracted. The energy area to be subtracted is determined by plotting the morning sounding and then plotting the afternoon maximum, assuming a dry adiabatic lapse rate relative to the original sounding. Then plot the next morning sounding. The energy area between the modified sounding of the first day and the morning sounding of the second day is the energy area to be subtracted. The second day maximum is plotted on the same chart with the morning sounding, thus modifying it to an afternoon sounding. Then the energy area is subtracted from this modified sounding to give the forecast minimum temperature for the third day. Again, this method has the disadvantage of the soundings not being normally made at the time of the minimum temperature. Too, the forecast value is not available at the time of the morning forecast, since it cannot be made until the maximum temperature for the day occurs.

FORECASTING SPECIAL SITUATIONS

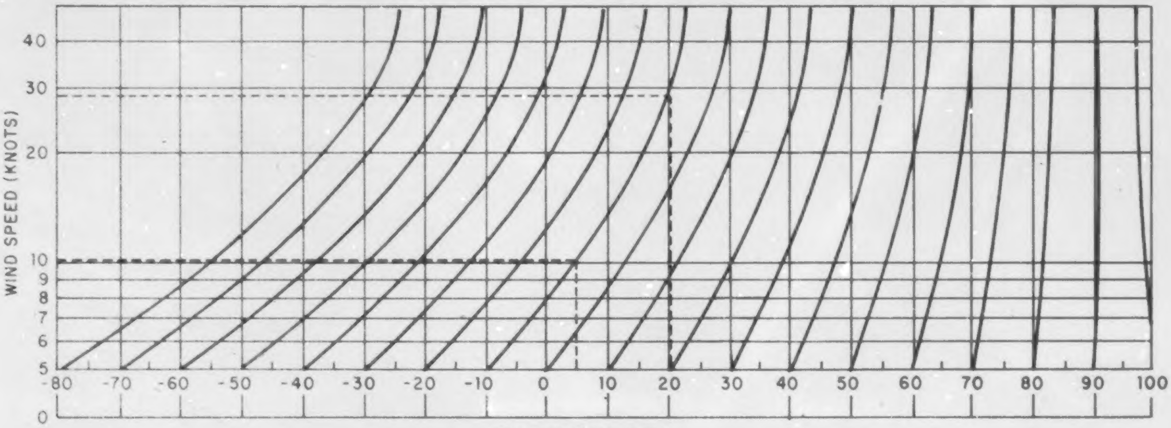
Cold Wave

A forecast of a cold wave gives warning of an impending severe change to much colder temperatures. In the United States it is defined as a net temperature drop of 20°F or more in 24 hours to a prescribed minimum that varies with geographical location and time of the year. Some of the prerequisites for a cold wave over the United States are continental polar air with temperature below average over west central Canada, movement of a low eastward from the Continental Divide which releases the cold wave, and large pressure tendencies on the order of 3 to 4 mb occurring behind the cold front. Aloft, a ridge of high pressure develops over the western part of the United States or just off the west coast, as a short wave trough moves into a long wave trough over the central part of the United States and deepens. An increase in intensity of the southwesterly flow over the eastern Pacific frequently precedes the building of the ridge. Frequently, retrogression of the long wave takes place. In any case, strong northerly to northwesterly flow is established aloft through a deep layer and sets the cP air in motion southward. When two polar outbreaks follow each other the second outbreak usually moves faster and overspreads the Central States. It also penetrates farther southward than the first cold wave. In such cases, the resistance of the southerly winds ahead of the second front is shallow. At middle and upper levels, winds remain west to northwest, and the long wave trough is situated near 80° west.

Most cold waves do not persist. Temperatures trend upward after about 48 hours. Sometimes, however, the upper ridge over the western part of the United States and the trough over the eastern part of the United States are quasi-stationary, and a large supply of very cold air remains in Canada. Then, we experience successive outbreaks with northwest steering that hold temperatures well below normal for as long as 2 weeks.

Heat Waves

In summer, heat wave forecasts furnish a warning that very unpleasant conditions are



INSTRUCTIONS, FROM THE INTERSECTION OF THE AIR TEMPERATURE AND THE WIND SPEED FOLLOW THE CURVED LINES TO THE BOTTOM AND READ THE WINDCHILL TEMPERATURE.

AG.757

Figure 10-32.–Windchill temperature graph.

impending. The definition of what is meant by a heat wave varies from place to place. For example, in the Chicago area a heat wave is said to exist when the temperature rises above 90°F on 3 successive days. In addition, there are many summer days which do not quite reach this requirement, but are highly unpleasant on account of humidity.

Heat waves develop over the midwestern and eastern part of the United States when a long wave trough stagnates over the Rockies or the Plains States and a long wave ridge lies over or just off the east coast. The belt of westerlies are centered far north in Canada. At the surface we observe a sluggish and poorly organized low-pressure system over the Great Plains or Rocky Mountains. Pressure usually is above normal over the South Atlantic and frequently the Middle Atlantic States. An exception occurs when the amplitude of the flow pattern aloft becomes very great. Then, several anticyclonic centers develop in the eastern ridge, both at upper levels and at the surface. Frequently, they are seen first at 500 mb. Between these meridionally arranged highs we see formation of east-west shear lines situated perhaps along latitudes 38° to 40°N. North of this line winds blow from the northeast and bring cool air from the Hudson Bay into the northern part of the United States. A general heat wave continues until the long wave train begins to move.

TEMPERATURE-HUMIDITY INDEX

In order to accurately express the comfort or discomfort caused by the air at various temperatures it is necessary to take into account the amount of moisture present. The National Weather Service uses the Temperature-Humidity Index to relate this.

The formula for computing the temperature-humidity index is given below:

$$\text{T.H.I.} = 0.4(t_d + t_w) + 15$$

where t_d is the dry-bulb temperature and t_w is the wet-bulb temperature. Both are in degrees Fahrenheit.

For example, suppose the temperature is 85°F and the wet-bulb temperature is 68°F. Substituting in the formula, we have:

$$\text{T.H.I.} = 0.4(85 + 68) + 15$$
$$\text{T.H.I.} = 0.4(153) + 15$$
$$\text{T.H.I.} = 76.2$$

The National Weather Service uses the values as follows: With a T.H.I. of 72, conditions are slightly uncomfortable; with a T.H.I. of 75, discomfort becomes more acute and most people would be using air conditioners, if available; with a T.H.I. of 79 and above, discomfort is general and air conditioning is highly desirable.

WINDCHILL TEMPERATURE

Weather office personnel will frequently be querried concerning windchill temperature by personnel engaged in outside tasks such as flight line or flight deck operations. The windchill temperature reflects the cooling power that the wind exerts on the air temperature.

Utilizing the wind speed and outside air temperature refer to figure 10-32 to find the windchill temperature.

From this temperature supervisors can determine the proper type clothing to be worn or if tasks can actually be performed.

CHAPTER 11

FORECASTING THUNDERSTORMS, FOG, TORNADOES, ICING, AND CONTRAILS

Forecasting of severe weather conditions and providing timely warnings for safety of aircraft and ship operations, as will as personnel safety, is a paramount responsibility of the senior Aerographer's Mate.

In this chapter we will discuss some of these phenomena and methods that may be utilized to forecast their occurrence.

THUNDERSTORMS

The thunderstorm represents one of the most formidable weather hazards in temperate and tropical zones. Though the effects of the thunderstorm tend to be localized, the turbulence, high winds, heavy rain, and occasionally, hail accompanying the thunderstorm are a definite threat to the safety of flight and to the security of naval installations. It is important that senior Aerographer's Mates be acquainted with the structure of thunderstorms and the types of weather associated with them, as well as being able to accurately predict their formation and movement.

Thunderstorm formation and movement were extensively covered in chapter 7 of AG 3 & 2; therefore in this chapter we will discuss, in more detail, the weather phenomena associated with thunderstorms and various forecasting methods.

THUNDERSTORM TURBULENCE AND WEATHER

Thunderstorms are characterized by turbulence, moderate to extreme updraft and downdrafts, hail, icing, lightning, precipitation, and under most severe conditions (in certain areas), tornadoes.

Turbulence (Drafts and Gusts)

Downdrafts and updrafts are vertical currents of air which are continuous over many thousand of feet of altitude, and are continuous over horizontal regions as large as a thunderstorm. Their speed is relatively constant as contrasted to gusts, which are smaller-scale discontinuities or variations in the windflow pattern extending over short vertical and horizontal distances. Gusts are primarily responsible for the bumpiness (turbulence) usually encountered in cumuliform clouds. A draft may be considered as a river flowing at a fairly constant rate, whereas a gust is comparable to an eddy or other type of random motion of water in a river.

Studies of the structure of the thunderstorm cell indicate that during the cumulus stage of development the updrafts may cover a horizontal area as large as 4 miles in diameter. In the cumulus stage the updraft in many cells extends from below the cloud base to the cloud top, a height greater than 25,000 feet. During the mature stage the updraft disappears from the lowest levels of the cloud although it continues in upper levels where it may exceed a height of 60,000 feet. These drafts are of considerable importance in flying because of the change in altitude which may occur when an aircraft flies through them.

In general, it has been found that the maximum number of high velocity gusts are found at altitudes of 5,000 to 10,000 feet below the top of the thunderstorm cloud, while the least severe turbulence is encountered, on the average near the base of the storm. This turbulence may at times also be classified as severe. The characteristic response of an aircraft intercepting a series of gusts is a number of sharp accelerations or "bumps" without a systematic change in altitude. The degree of bumpiness or turbulence experienced in flight is related to both the number of such abrupt changes encountered in a given distance and the strength of the individual changes. All date in this section are based on information from the Thunderstorm Project in 1947.

Hail

Hail is regarded as one of the worst hazards of thunderstorm flying. It usually occurs during the mature stage of cells having an updraft of more than average intensity, and is found with the greatest frequency between 10,000- and 15,000-ft levels. As a rule, the larger the storm the more likely it is to have hail.

Although encounters by aircraft with large hail are not too common, hail rather than one-half or three-fourths inch can damage an aircraft in a very few seconds. The general conclusion regarding hail is that most mid-latitude storms contain hail sometime during their life cycle with most hail occurring during the mature stage. In subtropical and tropical thunderstorms, hail seldom reaches the ground. It is generally believed that these thunderstorms contain less hail aloft than do mid-latitude storms.

Rain

Thunderstorms contain considerable quantities of moisture which may or may not be falling to the ground as rain. These water droplets may be suspended in, or moving with, the updrafts. Rain is encountered below the freezing level in almost all penetrations of fully developed thunderstorms. Above the freezing level, however, there is a sharp decline in the frequency of rain.

There seems to be a definite correlation between turbulence and precipitation. The intensity of turbulence, in most cases, varies directly with the intensity of precipitation. This relationship indicates that most rain and snow in thunderstorms is held aloft by updrafts.

Icing

Where the free-air temperatures are at or below freezing, icing should be expected in flights through thunderstorms. In general, icing is associated with temperatures from 0° to –20°C. Most severe icing occurs from 0°C to –10°C. The heaviest icing conditions usually occur in that region above the freezing level where the cloud droplets have not yet turned to ice crystals. When the thunderstorm is in the cumulus stage severe icing may occur at any point above the freezing level. However, because of the formation of ice crystals at high levels and the removal of liquid water by precipitation, icing conditions are usually somewhat less in the mature and dissipating stage.

THUNDERSTORM ELECTRICITY AND LIGHTNING

The thunderstorm changes the normal electric field, in which the earth is negative with respect to the air above it, by making the upper portion of the thunderstorm cloud positive and the lower part negative. This negative charge then induces a positive charge on the ground. The distribution of the electric charges in a typical thunderstorm is shown in figure 11-1. The lightning first occurs between the upper positive charge area and the negative charge area immediately below it. Lightning discharges are considered to occur most frequently in the area bracketed roughly by the 32°F and the 15°F temperature levels. However, this does not mean that all discharges are confined to this region, for as the thunderstorm develops, lightning discharges may occur in other areas, and from cloud to cloud, as well as cloud to ground. Lightning can do considerable damage to aircraft, especially to radio equipment.

THUNDERSTORMS IN RELATION TO ENVIRONMENTAL WIND FIELD

During all stages of a cell, air is being brought into the cloud through the sides of the cloud. This process is known as entrainment. A cell

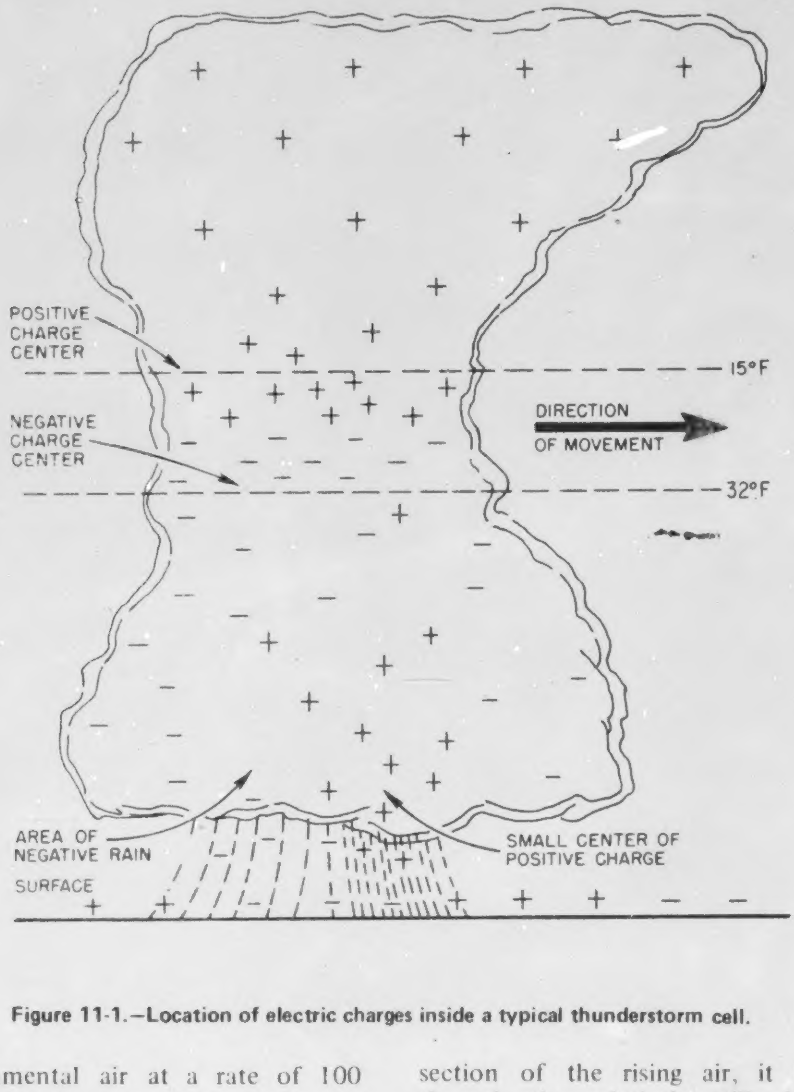


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Figure 11-1.—Location of electric charges inside a typical thunderstorm cell.

entrains environmental air at a rate of 100 percent per 500 mb, that is it doubles its mass in an ascent of 500 mb. The factor of entrainment is important in establishing a lapse rate within the cloud which is greater than the moist adiabat and in maintaining the downdraft.

When there is a marked increase with height in the horizontal wind speed, the mature stage of the cell may be prolonged. In addition, the increasing speed of the wind with height produces considerable tilt to the updraft of the cell, and in fact, to the visible cloud itself. Thus, the falling precipitation passes through only a small section of the rising air, it falls thereafter through the relatively still air next to the updraft, perhaps even outside the cell boundary. Therefore, since the drag of the falling water is not imposed on the rising air currents within the thunderstorm cell, the updraft can continue until its source of energy is exhausted. Tilting of the thunderstorm explains why hail is sometimes encountered in a cloudless area just ahead of the storm.

RADAR DETECTION

Radar, either surface or airborne, is the best aid in detecting thunderstorms, charting their movement, or selecting an area for feasible penetration. A thunderstorm's size, direction of movement, shape and height, as well as other significant features, can be determined from a radar presentation. Radarscope interpretation is discussed in chapter 16 of this training manual.

THUNDERSTORM FLIGHT HAZARDS

Thunderstorms are virtually weather factories in that the pilot flying into one can expect to encounter great variations in the weather; some of them hazardous. Thunderstorms are often accompanied by extreme fluctuations in ceiling and visibility. Every thunderstorm has turbulence, sustained updrafts and downdrafts, precipitation, and lightning. Icing conditions, though quite localized, are quite common in thunderstorms, and many contain hail. The flight conditions listed below are generally representative of many (but not necessarily all) thunderstorms.

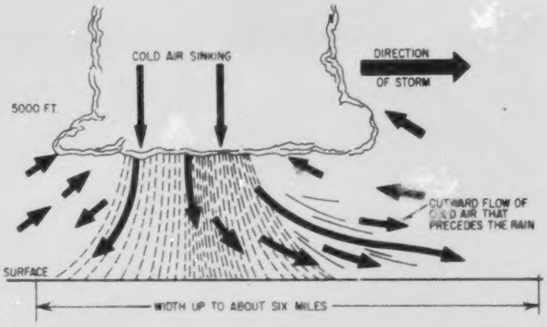
1. The chance of severe or extreme turbulence within thunderstorms is greatest at higher altitudes, with most cases of severe and extreme turbulence about 8,000 to 15,000 feet above terrain. The least turbulence may be expected when flying at or just below the base of the main thunderstorm cloud. (The latter rule would not be true over rough terrain or in mountainous areas where strong eddy currents produced by strong surface winds would extend the turbulence up to a higher level.)
2. The heaviest turbulence is closely associated with the areas of heaviest rain.
3. The strongest updrafts are found at heights of about 10,000 feet or more above the terrain; in extreme cases, updrafts in excess of 65 feet per second occur. Downdrafts are less severe, but downdrafts on the order of 20 feet per second are quite common.
4. The probability of lightning strikes occurring is greatest near or slightly above the freezing level.

Because of the potential hazards of thunderstorm flying, it is obviously nothing short of folly for pilots to attempt to fly in thunderstorms, unless operationally necessary.

THUNDERSTORM SURFACE PHENOMENA

The rapid change in wind direction and speed immediately prior to a thunderstorm passage is a significant surface hazard associated with thunderstorm activity. The strong winds which accompany thunderstorm passages are the result of the horizontal spreading out of downdraft currents from within the storm as they approach the surface of the earth.

Figure 11-2 shows the nature of the wind outflow and indicates how it is formed from the settling dome of cold air which accompanies the rain core during the mature stage of the thunderstorm. The arrival of this outflow results in a radical and abrupt change in the wind speed and direction. It is an important consideration in the landing and taking off of aircraft in advance of the arrival of a thunderstorm.


COLD AIR SINKING
DIRECTION OF STORM
5000 FT.
OUTWARD FLOW OF COLD AIR THAT PRECEDES THE RAIN
SURFACE
WIDTH UP TO ABOUT SIX MILES

AG.608

Figure 11-2.—Cold dome of air beneath a thunderstorm cell in the mature stage. Arrows represent deviation of windflow. Dashed lines indicate rainfall.

Wind speeds at the leading edge of the thunderstorm are ordinarily far greater than those at the trailing edge. The initial wind surge observed at the surface is known as the first gust. The speed of the first gust is normally the highest recorded during the storm passage and may vary as much as 180 degrees in direction from the surface wind direction which previously existed. The mass of cooled air spreads out from downdrafts of neighboring thunderstorms (especially in squall lines) and often becomes organized into a small, high-pressure area called a bubble-high or meso-high, which persists for some time as an entity that can sometimes be seen on the surface map. These highs may be a mechanism for controlling the direction in which new cells form.

The speed of the thunderstorm wind depends upon a number of factors, but local surface winds reaching up to 50 to 75 miles per hour for a short time are not uncommon. Because it can extend for several miles in advance of the thunderstorm itself, the thunderstorm wind is a highly important consideration for pilots preparing to land or take off in advance of a storm's arrival. Also, many thunderstorm winds are strong enough to do considerable structural damage and capable of overturning or otherwise damaging even medium sized aircraft that are parked and are not adequately secured.

The outflow of air ahead of the thunderstorm sets up considerable low level turbulence. Over relatively even ground, most of the important turbulence associated with the outrush of air is within a few hundred feet of the ground, but it extends to progressively higher levels as the roughness of the terrain increases.

THUNDERSTORM ALTIMETRY

During the passage of a thunderstorm, rapid and marked surface pressure variations generally occur. These variations usually occur in a particular sequence characterized by the following.

1. An abrupt fall in pressure as the storm approaches.
2. An abrupt rise in pressure associated with rain showers as the storm moves overhead (often associated with the first gust).
3. A gradual return to normal pressure as the storm moves on and the rain ceases.

Such pressure changes may result in significant altitude errors on landing.

Of greater concern to the pilot are pressure readings which are too high. If a pilot used an altimeter setting given to him during the maximum pressure and then landed after the pressure had fallen, he would have found that his altimeter still read 60 feet or more above the true altitude after he was on the ground.

Here is where you, as a section leader, or office supervisor, can make certain that timely and accurate altimeter settings are furnished to the tower for transmission to pilots during thunderstorm conditions. If the data are old and inaccurate a crash could result.

THUNDERSTORM FORECASTING

The standard method of forecasting air mass thunderstorms has long consisted primarily of an analysis of radiosonde data with particular emphasis on the so-called positive areas, in which the temperature of a lifted parcel exceeds the temperature of its surroundings and, hence, being less dense, must rise, as a cork in water.

Experienced forecasters are aware that this method, though basic, is also crude and at times wholly inaccurate. Many times conditions are favorable for thunderstorm development with a large positive energy area showing up on the sounding, with no ensuing thunderstorm activity whatever. At other times thunderstorms occur when they should not and where they should not be. Clearly, factors other than the one of instability are important, and at times of overriding importance.

During recent years forecasters have come to recognize this need for better forecasting methods. A number of methods have been developed, but many of these are too complicated or detailed to include in this training course. Covered in this section are the forecasting of convective clouds by the parcel method, with conditions necessary for thunderstorm development; the development of thunderstorms by mechanical lifting (orographic or frontal); a brief discussion of the importance of the slice method; and a method for the prediction of these storms which enables the average forecaster to arrive at a fairly accurate, reasonably objective forecast of these storms. Credit is given to the Academic Press for permission to use this method from Weather Forecasting for Aeronautics, by J. J. George and Associates, Academic Press, 1960.

For a more detailed discussion of the determination of instability, stability, the convective condensation level (CCL), the level of free convection (LFC), the lifting condensation level (LCL), and the slice method, refer to chapter 3 of this training manual.

THE PARCEL METHOD

Nearly all of the procedures routinely used to evaluate and analyze the stability of the atmosphere are manipulations of the parcel method. The temperature of a minute parcel of air is assumed to change adiabatically as the parcel is displaced a small distance vertically from its original position. If, after vertical displacement, the parcel has a higher vitural temperature than the surrounding atmosphere, the parcel is subjected to a positive buoyancy force and will be further accelerated upwards; conversely, if its virtual temperature has become lower than that of the surrounding air, the parcel will be denser than its environment, thus subjected to a negative buoyancy force, is retarded, and eventually returns to its initial or equilibrium position.

Formation of Clouds by Heating From Below

The first step is to determine the convection temperature or the surface temperature that must be reached to start the formation of convection clouds by solar heating of the surface air layer. The procedure is to first determine the CCL on the plotted sounding and, from the CCL point on the T curve of the sounding, proceed downward along the dry adiabat to the surface-pressure isobar. The temperature read at this intersection is the convection temperature.

Figure 11-3 shows an illustration of forecasting afternoon convective cloudiness from a plotted sounding. The dewpoint curve was not plotted to avoid confusion. The dashed line with arrowheads indicates the path the parcel of air would follow under these conditions. You can see that the sounding was modified at various times during the day.

To determine the possibility of thunderstorms by the use of this method and from an analysis of the sounding, the following conditions must exist:

1. Sufficient heating must occur.
2. The positive area must exceed the negative area. The greater the excess, the greater the possibility of thunderstorms.
3. The parcel must rise to the ice crystal level. Generally, this level should be −10°C and below.
4. There must be sufficient moisture in the lower troposphere. This has been found to be the most important single factor in thunderstorm formation.
5. Climatic and seasonal conditions should be favorable.
6. Weak inversions (or none at all) should be present in the lower levels.
7. An approximate height of the cloud top may be determined by assuming that the top of the cloud will extend beyond the top of the positive area by a distance equal to one-third of the height of the positive area.

Formation of Clouds by Mechanical Lifting

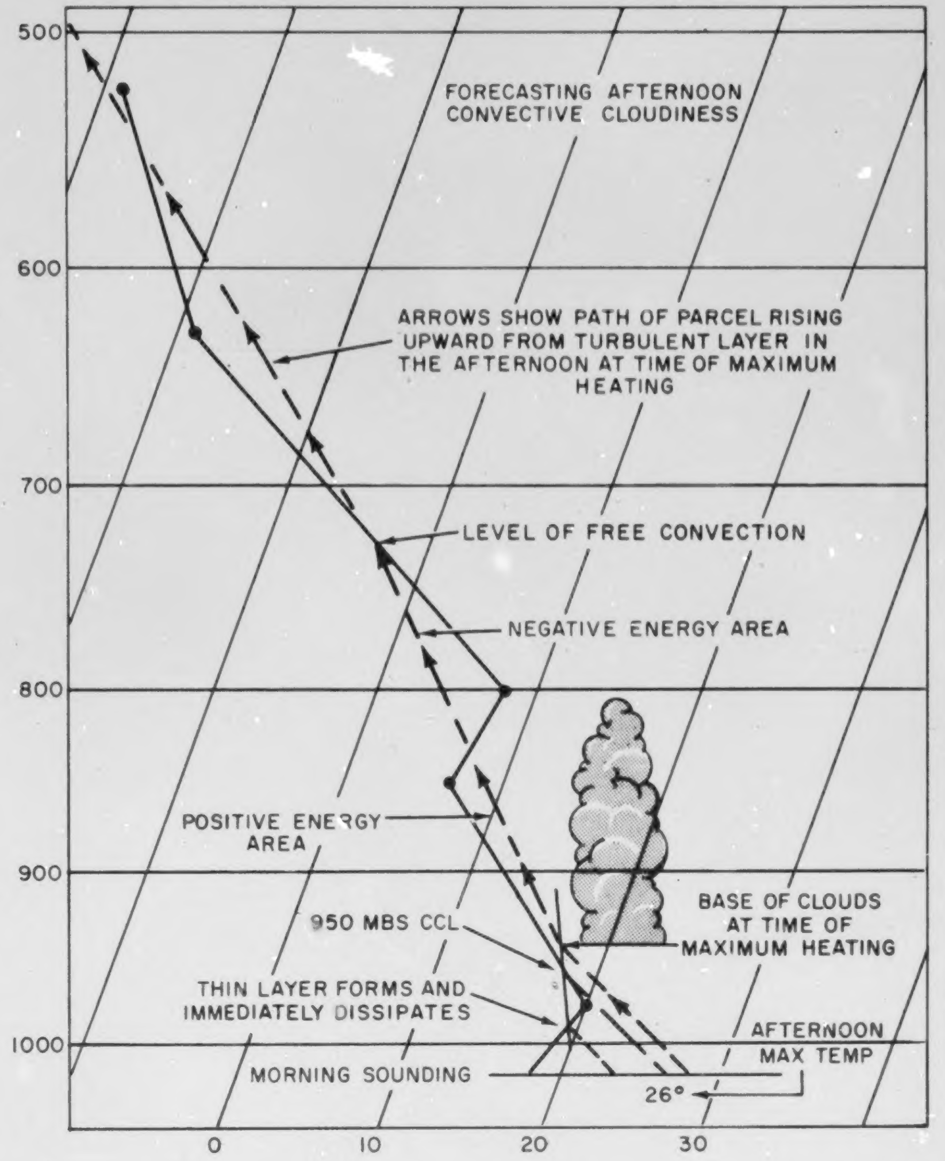
When using this method, it is assumed that the type lifting will be either orographic or frontal. Here we are working with the LCL and the LFC.

The LCL is the height at which a parcel of air becomes saturated when it is lifted dry adiabatically. The LCL for a surface parcel is always found at or below the CCL. The LFC is the height at which parcel of air lifted dry adiabatically until saturated and saturation adiabatically thereafter would first become warmer than the surrounding air. The parcel will then continue to rise freely above this level until it becomes colder than the surrounding air.

Figure 11-4 shows the formation of clouds due to mechanical lifting. This figure shows the formation of a stratified layer of clouds above the LCL to the LFC. At the LFC and above, the clouds would be turbulent. The tops of the clouds extend beyond the top of the positive area due to overshooting just as in the case of clouds formed due to heating.

To determine the possibility of thunderstorms from this method, the following conditions should be met:

1. The positive area must exceed the negative area; the greater the excess, the greater the possibility of thunderstorms.
2. There must be sufficient lifting to lift the parcel to the LFC. The frontal slope or the Orographic barriers can be used to determine how much lifting can be expected.
3. The parcel must reach the ice crystal level.


500
600
700
800
900
1000
0
10
20
30
FORECASTING AFTERNOON CONVECTIVE CLOUDINESS
ARROWS SHOW PATH OF PARCEL RISING UPWARD FROM TURBULENT LAYER IN THE AFTERNOON AT TIME OF MAXIMUM HEATING
LEVEL OF FREE CONVECTION
NEGATIVE ENERGY AREA
POSITIVE ENERGY AREA
950 MBS CCL
THIN LAYER FORMS AND IMMEDIATELY DISSIPATES
BASE OF CLOUDS AT TIME OF MAXIMUM HEATING
AFTERNOON MAX TEMP
MORNING SOUNDING
26°

AG.609

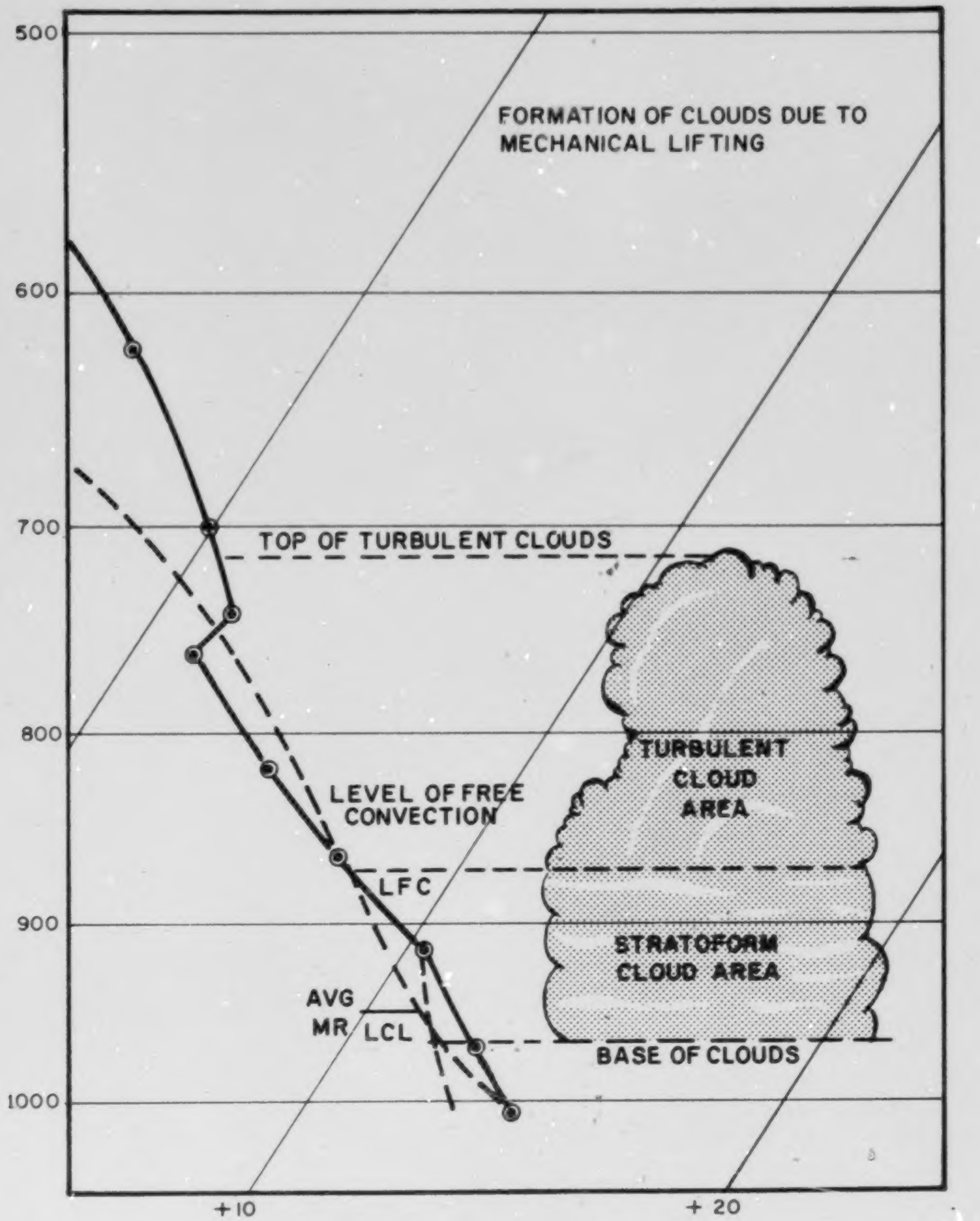
Figure 11-3.—Forecasting afternoon convective cloudiness.

4. Even though the positive area does not exceed the negative area, cloudiness occurs after the parcel passes the LCL and precipitation may occur after the parcel passes the ice crystal level.

One main advantage of this method is that it can be done quickly and with relative ease. A major disadvantage is that it assumes that the parcel does not change its environment and that it overestimates or underestimates the stability and instability conditions.

INSTABILITY INDICATIONS FROM THE WET-BULB CURVE

A layer of the atmosphere is potentially unstable if the potential wet-bulb temperature


500
600
700
800
900
1000
FORMATION OF CLOUDS DUE TO MECHANICAL LIFTING
TOP OF TURBULENT CLOUDS
TURBULENT CLOUD AREA
LEVEL OF FREE CONVECTION
LFC
STRATOFORM CLOUD AREA
AVG MR
LCL
BASE OF CLOUDS
+10
+ 20

AG.610

Figure 11-4.—Formation of clouds due to mechanical lifting.

decreases with altitude. Potential instability refers to a layer that is lifted as a whole. Some meteorologists use the term "convective" vice "potential;" however, potential causes less configuration regarding the lifting process involved. The wet-bulb temperature may be found by lifting each individual point on the sounding dry adiabatically to saturation and then back to its original level moist adiabatically. By connecting the points on a sounding, a wet-bulb curve can be constructed.

If the wet-bulb curve slopes to the right with increasing altitude, the potential wet-bulb temperature increases with height and the layer is potentially stable. If it slopes to the left with increasing height, more than the saturation adiabats, the layer is potentially unstable. If none of the potential curves intersect the sounding, thunderstorms are not likely to occur.

BAILEY GRAPH METHOD

Credit is given to the Academic Press, New York City, for permission to use this method from Weather Forecasting for Aeronautics, by J. J. George and Associates. It is mainly for use in the Eastern United States.

The Bailey Graph method is a reasonably simple method for forecasting air mass thunderstorms, but does not require prediction of short range change in the vertical distribution of temperature and moisture. This method is best used with the 1200Z sounding.

The first step is to eliminate those areas whose soundings disclosed moisture inadequacies. This is done by means of the following six steps:

1. Dewpoint depression 13°C or more at any level from 850 through 700 mb.
2. Dewpoint depression sum of 28° or more at 700- and 600-mb levels.
3. Dry or cool advection at low levels.
4. Surface dewpoint 60°F or less at 0730 local with no substantial increase expected before early afternoon.
5. Lapse rate 21°C or less from 850 to 500 mb.
6. A freezing level below 12,000 feet in unstable cyclonic flow produces only light showers.

The two parameters used in making the forecast, after eliminating all soundings meeting one or more of the above six conditions, were simply (1) the lapse rate between 850 and 500 mb (the difference in temperature between these two levels–for example, if the temperature at 850 mb was 15°C and at 500 mb it was -10°C, the difference would equal 25°C); and (2) the sum of the dewpoint depressions at 700 and 600 mb in degrees C. These two figures are used as arguments for the graph in figure 11-5.

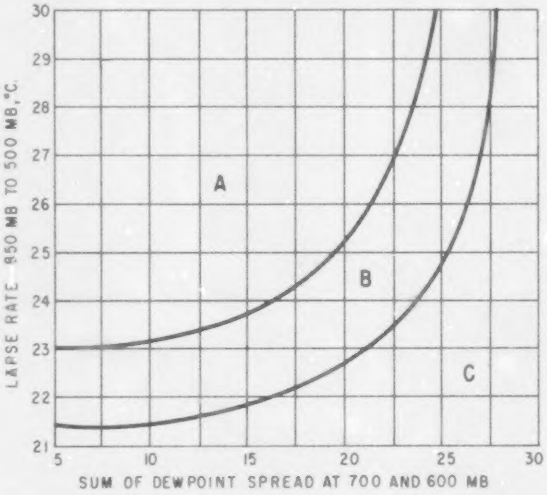


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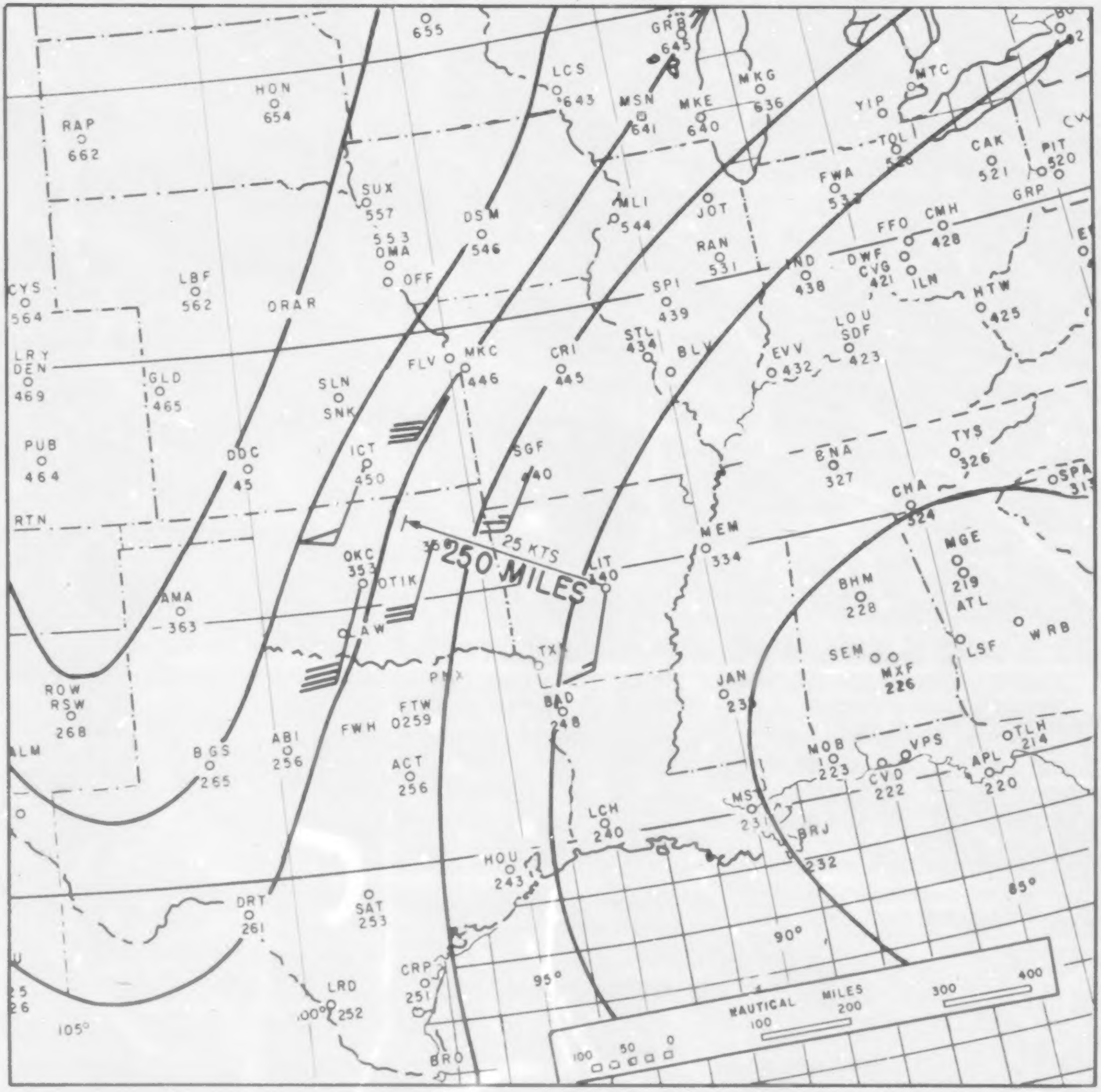
Figure 11-5.—Local area thunderstorm graph. Area "A" is isolated thunderstorms, with a 12 to 1 chance of at least one rain gage in dense network receiving rain. Area "B" is scattered thunderstorms, with a 4 to 1 chance of reported rain. Area "C" is no rain.

One further condition for the development of thunderstorms was found to be the absence of large anticyclonic wind shear which is measured at 850 mb. This condition is that no horizontal shear at this level may exceed 20 knots in 250 miles measured toward low pressure from the sounding station. Figure 11-6 illustrates how this measurement is made.

STABILITY INDEXES AS AN INDICATION OF INSTABILITY

The overall stability or instability of a sounding is sometimes conveniently expressed in the


250 MILES
25 KTS
RAP 662
HON 654
655
GRB 645
LCS 643
MSN 641
MKE 640
MKG 636
YIP
MTC
TOL 526
CAK 521
PIT 520
SUX 557
553 OMA
OFF
DSM 546
MLI 544
JOT
FWA 533
FFO
CMH 428
GRP
RAN 531
IND 438
DWF
CVG 421
ILN
HTW 425
CYS 564
LBF 562
ORAR
SPI 439
STL 434
LOU SDF 423
EVV 432
BLV
LRY DEN 469
GLD 465
FLV
MKC 446
CRI 445
SLN SNK
PUB 464
DDC 45
ICT 450
SGF 440
BNA 327
TYS 326
OSPA
CHA 324
RTN
MEM 334
LIT 340
MGE 219
ATL
BHM 228
OKC 353
TIK
AMA 363
LAW
TXK
PNX
WRB
LSF
SEM
MXF 226
JAN 23
ROW RSW 268
FTW 0259
FWH
BAD 248
ABI 256
ALM
BGS 265
ACT 256
MOB 223
VPS
CVD 222
TLH 214
APL 220
LCH 240
MSY 231
BRJ 232
HOU 243
DRT 261
SAT 253
CRP 251
LRD 252
BRO
105°
100°
95°
90°
85°
NAUTICAL MILES
100 50 0
100 200 300 400

AG.612

Figure 11-6.—Example of anticyclonic shear and curvature at 850 mb preventing thunderstorms, with otherwise favorable air mass conditions present in the Little Rock area.

form of a single numerical value called the stability index. Such indexes have been introduced mainly as aids in connection with particular forecasting techniques or studies. Most of the indexes take the form of a difference in temperature, dewpoint, wet-bulb temperature, or potential temperature in height or pressure between two arbitrarily chosen surfaces. These indexes are generally useful only when combined, either objectively or subjectively, with other data and synoptic considerations. Used alone, they are less valuable than when plotted on stability index charts and analyzed for large areas. In this respect they have the value of alerting the forecaster to those soundings, routes, or areas which should be more closely examined by other procedures.

There are a number of methods that may be utilized for determining stability or instability; among these are the Showalter Index (SI), the Lifted Index (LI), the Fawbush-Miller Stability Index (FMI), and the Martin Index (MI).

The National Weather Service currently employs the Lifted Index method for producing their facsimile products. All current methods are discussed in detail in NAVAIR 50-1P-5, Use of the Skew T, Log P Diagram in Analysis and Forecasting; however, in this manual we will discuss the Showalter Index method only.

Showalter Index (SI)

This is the most widely used of the various types of indexes.

The procedure for computing the Showalter Stability Index is explained in the following section from the illustration in figure 11-7.

Step 1. From the 850-mb temperature (T), draw a line parallel to the dry adiabat upward until it intersects the saturation mixing ratio line for the dewpoint temperature at 850 mb. This is called the LCL on this diagram. A mountain station would have to use some higher level.

Step 2. From the LCL, draw a line parallel to the saturation adiabat upward to 500 mb. Let the temperature at this intersection point at 500 mb be called T′.

Step 3. Algebraically, subtract T′ from the 500-mb temperature. The value of the remainder, including its algebraic sign, is the value of the Showalter Index. In figure 11-7, T′ = -25°C and T = 22°C; the Showalter Index is therefore +3°. This index is positive when T′ lies to the left of the T curve. Positive index values imply greater stability of the sounding.

For forecasting purposes in the United States, the significance of the index values to forecasting is as follows:

1. When the index is +3° or less, showers are probable and some thunderstorms may be expected in the area.
2. The chance of thunderstorms increases rapidly for index values in the range of +1° to -2°.
3. Index values of -3° or less are associated with severe thunderstorms.
4. When the value of the index is below -6°, the forecaster should consider the possibility of tornado occurrence. However, the forecasting value of all index categories must, in each case, be evaluated in the light of the moisture content of the air and of other synoptic conditions.

FORECASTING MOVEMENT OF THUNDERSTORMS

Radar can be an invaluable aid in determining the speed and the direction of movement of the thunderstorm. Sometimes it is desirable and necessary to estimate the movement from winds aloft. There is no completely reliable relationship between speed of the winds or direction of the winds aloft in forecasting thunderstorm movement. However, one study reveals that there is a marked tendency for large convective rainstorms to move to the right of the wind direction in the mean cloud layer from 850 to 500 mb (or the 700-mb wind) with a systematic deviation of about 25 degrees to the right of the flow. There appears to be little correlation between wind speed and speed of movement at any level, although the same study mentioned above revealed that 82 percent of the storms move within plus or minus 10 knots of a mean speed of 32 knots.

FORECASTING MAXIMUM GUSTS WITH NONFRONTAL THUNDERSTORMS

It should be remembered that maximum gusts associated with thunderstorms occur over a very small portion of the area in which the thunderstorm exists and usually occur immediately prior to the storm's passage. Nevertheless, the possibility of damage to aircraft and installations on the surface is so great that every available means should be used to make the best and most accurate forecast possible to forewarn the agencies concerned.

Estimation of Gusts From Climatology and Storm Intensity

The Aerographer's Mate is aware that the season of the year and the location of the

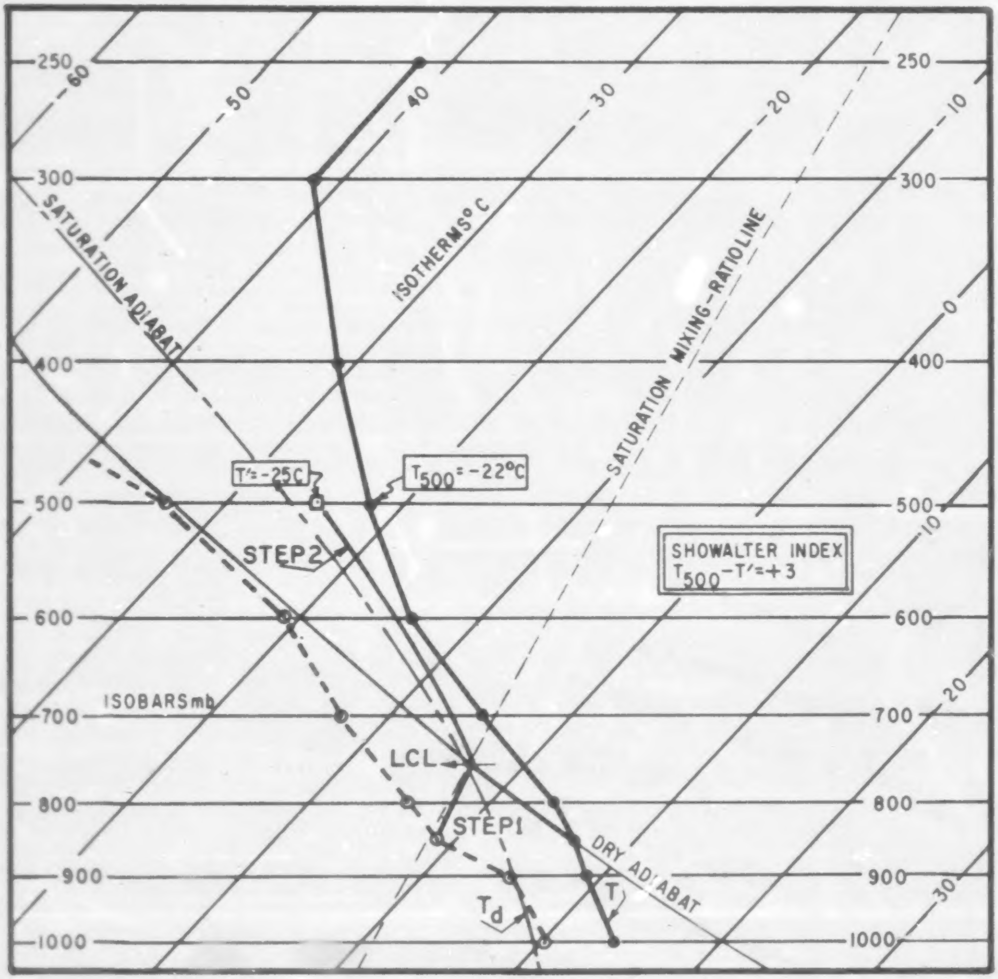


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Figure 11-7.—Computation of the Showalter Stability Index.

station have a great bearing upon the maximum winds to be expected. Certain areas of the United States, and the world, have a history of severe thunderstorm occurrence with ensuing strong winds during the most favorable seasons of the year. For this reason you should have the thunderstorm climatology for your station, as well as for the general area, available as to time of occurrence, season of occurrence, and the associated conditions attending them.

The storm's intensity and those reports from neighboring or nearby stations can give you a good indication of what conditions to expect at your terminal.

Forecasting Peak Wind Gusts Using USAF Method

There are two basic methods for forecasting maximum wind gusts of convective origin outlined in AWS Technical Report 200 (Rev). One involves the use of the Dry Stability Index (T_1) and the other downrush temperature subtracted from the dry-bulb temperature (T_2). These methods will be discussed in this section.

DETERMINATION OF T_1.–T_1 is found in one of two ways:

1. If the sounding has an inversion, the moist adiabat is followed from the warmest point of the inversion to 600 millibars. The temperature difference between the intersection of the moist adiabat at the 600 mb isobar and the temperature of the dry bulb at 600 mb is T_1. The inversion (top) point should be within 150 or 200 mb of the surface and must not be susceptible to becoming wiped out by surface convection.

2. If no inversion appears on the sounding or if the inversion is relatively high (more than 200 mb above the surface), a different method is used to find T_1. The maximum temperature at the surface is forecast in the usual manner. A moist adiabat is projected from the maximum temperature to the 600 mb level. The temperature difference between the intersection of the moist adiabat and the 600 mb surface and the dry-bulb temperature at the 600 mb level is T_1.

After T_1 has been determined refer to Table 11-1 for determination of the maximum gust speed. To further add to the reliability and accuracy of the forecast, one-third of the mean wind speed expected in the lower 5,000 feet above the ground level should be added to the value obtained from Table 11-1.

DETERMINATION OF T_2.–T_2 is found by first locating the 0°C isotherm on the wet-bulb curve. A moist adiabat through that point is followed down to the surface and the temperature at that point is recorded. This temperature is subtracted from the dry-bulb temperature (or the forecast free air temperature) giving the value of T_2.

Using the value of T_2 refer to figure 11-8. This graph allows you to arrive at the probable minimum, mean, and maximum gusts.

DETERMINATION OF GUST DIRECTION.–For the direction of the maximum gust, the mean wind direction in the layer between 10,000 and 14,000 feet above the terrain is used.

EXAMPLES OF FORECASTING METHODS.–Figure 11-9 shows a typical sounding plotted on the Skew T diagram for forecasting maximum gust speeds.

The following procedure may be followed:

1. A moist adiabat is projected from the warmest point of the inversion to the 600 mb surface, and the temperature at that intersection is 0.8°C.

2. The dry-bulb temperature at 600 mb is −7.8°C, so that T_1 is about 9°C.

3. Entering Table 11-1, the value of V′ is found to be 35 knots. To this value add one-third of the mean wind speed in the layer from the surface to 5,000 feet to obtain the maximum peak gust.

Table 11-1.–Use of T_1 for maximum wind gusts.

T_1 Values in °C	Maximum Gust Speed (V′)	T_1 Values in °C	Maximum Gust Speed (V′)
3	17	14	47
4	20	15	49
5	23	16	51
6	26	17	53
7	29	18	55
8	32	19	57
9	35	20	58
10	37	21	60
11	39	22	61
12	41	23	63
13	45	24	64
		25	65

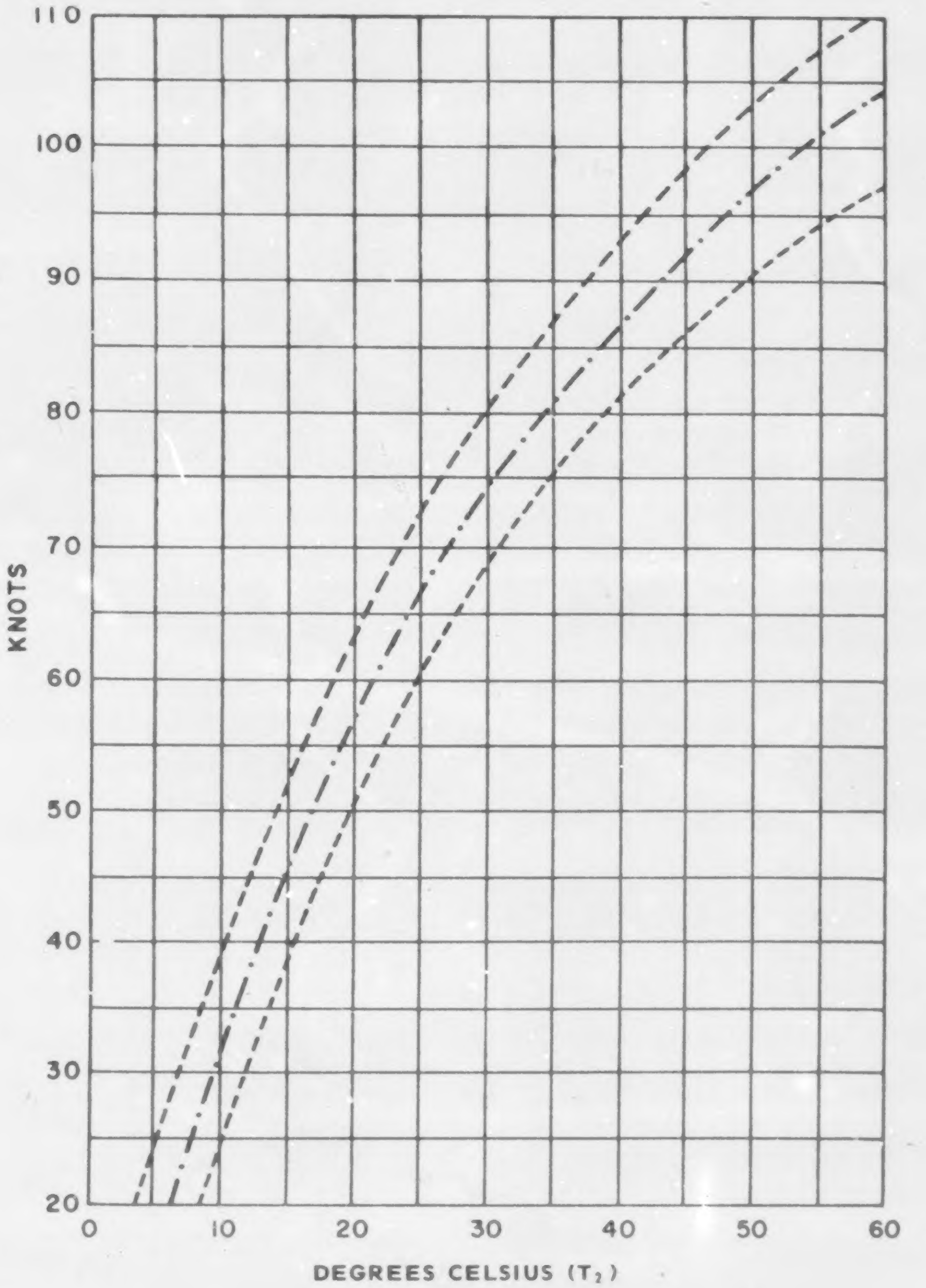

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Figure 11-8.—Probable maximum gusts using the T_2 method.

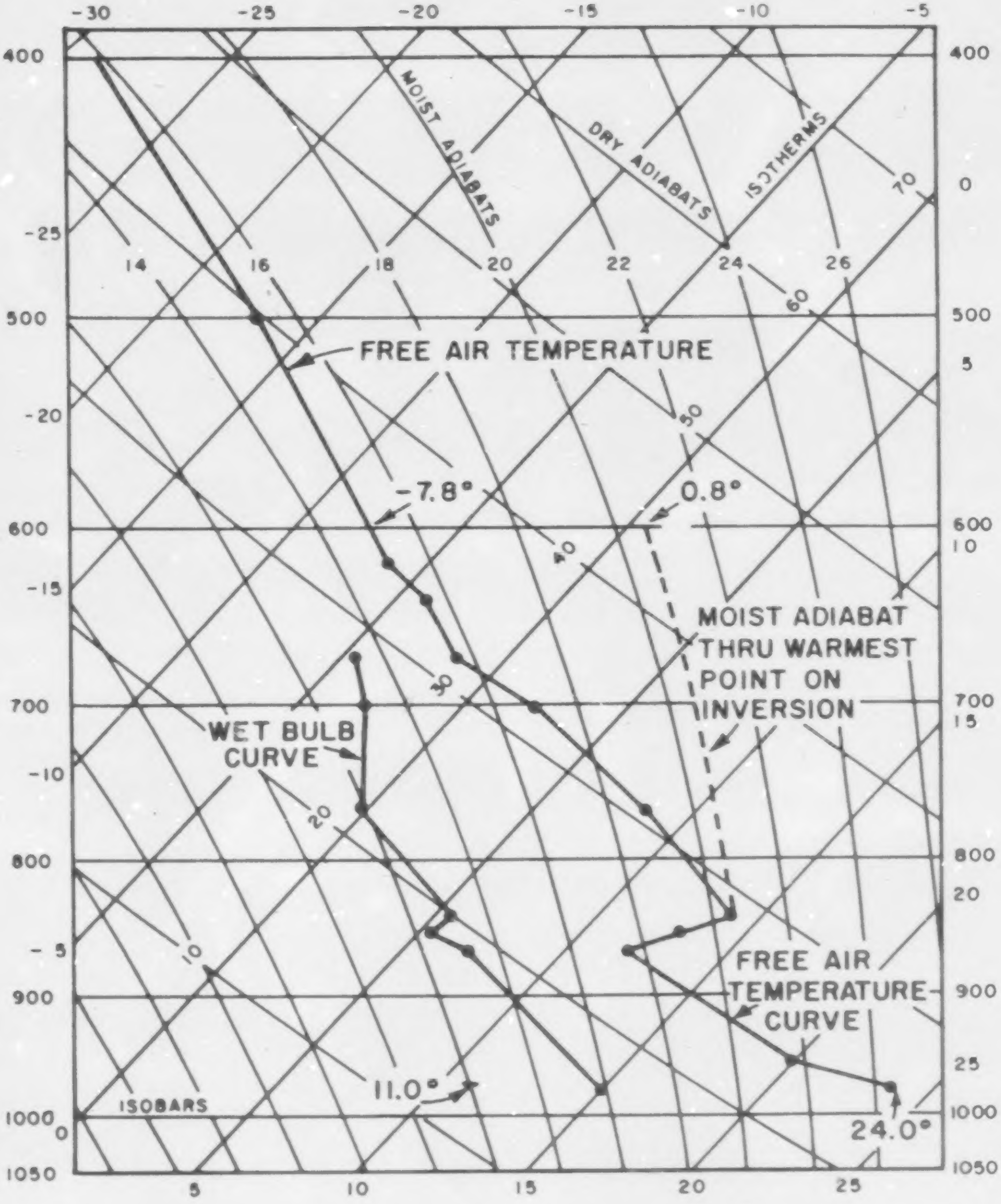

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Figure 11-9.—Example of sounding for wind gust forecast.

4. The forecast wind direction of the maximum gust is the same as the mean wind direction from 10,000 to 14,000 feet above the terrain.

Figure 11-10 is used to illustrate the method of forecasting maximum wind gusts using the T_2 method.

2. Since the surface dry-bulb temperature is 27°C the value of T_2 is 15°C.

3. Entering figure 11-8 at T_2 = 15 the probable minimum wind speed is 38 knots; the mean speed is 45 knots; and probable maximum is 52 knots.

4. Wind direction is obtained in the same manner as the preceding method.

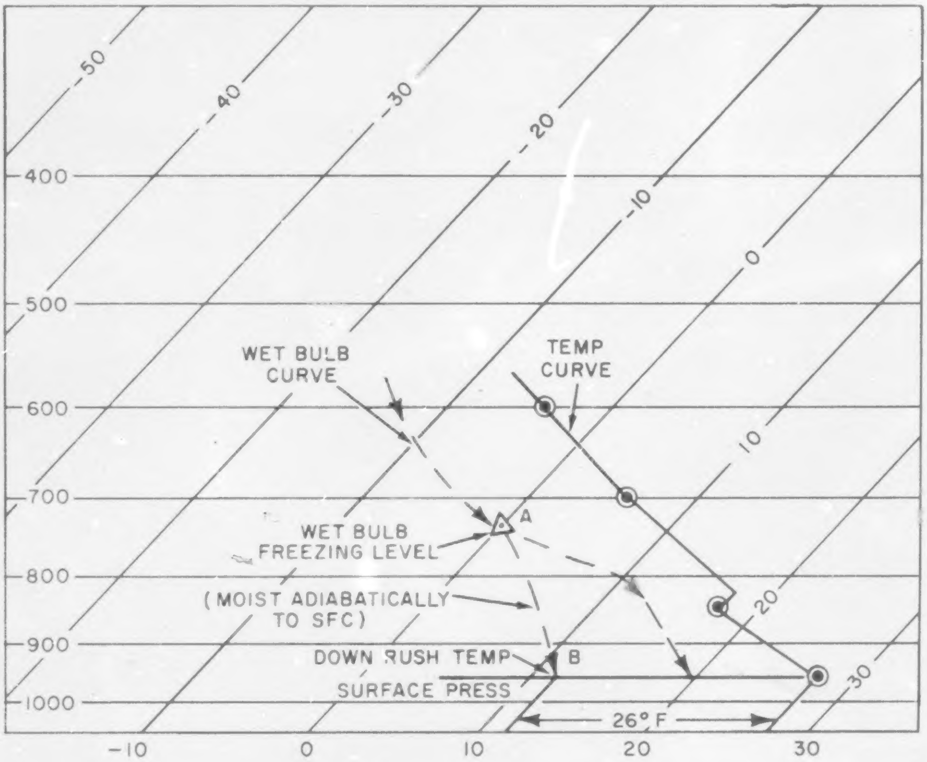

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Figure 11-10.—Example of sounding for forecasting maximum wind gusts using downrush temperature.

The following steps show the correct procedure for determining the value of T_2:

1. A moist adiabat is projected downward from the wet bulb freezing level to the surface and the temperature at the intersection is found to be 12°C.

FORECASTING HAIL

Hail, like the maximum wind gusts in thunderstorms, usually takes place in a narrow shaft seldom wider than a mile or two and usually less than a mile wide. The occurrence of hail in thunderstorms was discussed earlier in this chapter. However, a few additional facts concerning the occurrence and frequency of hail in flight should be discussed at this point.

1. Since hail is normally associated with thunderstorms, the season of the maximum occurrence of hail is coincident with the season of maximum occurrence of thunderstorms.

2. When the storm is large and well developed, an assumption should be made that it contains hail.

3. Encounters of hail below 10,000 feet showed the hail distribution equally divided between the clear air alongside the thunderstorm, in the rain area underneath the storm, and within the thunderstorm itself.

4. From 10,000 to 20,000 feet, the percentages ranged from 40 percent in the clear air alongside the storm to 60 percent in the storm, with 82 percent of the encounters outside the storm under the overhanging cloud.

5. Above 20,000 feet, the percentages reflected 80 percent of the hail was encountered in the storm with 20 percent in the clear air beneath the anvil or other cloud extending from the storm.

Climatology is of vital importance in predicting the hail occurrence, as well as its size. Good estimations of the size of hail can be gleaned from reports of the storm passage over nearby upstream stations. Here too, modifying influences must be taken into account.

Hail Frequency as Related to Storm Intensity and Height

Fifty percent hail occurrency may be expected in storms exceeding 46,000 feet, based on radar echo height alone. With a maximum echo height of 52,000 feet, a 67 percent hail frequency can be expected; and only 33 percent at 35,000 feet. Mean echo heights are 42,000 feet for hail and 36,000 feet for rain.

A Yes-No Hail Forecasting Technique

Aside from the fact that hail occurs with thunderstorms both inside and outside the storm and our knowledge of the relationship of hail to this type of storm, very little information is available as to forecasting of the actual occurrence of hail. The technique presented in this section of the chapter is an objective method of hail forecasting, using the parameters of the ratio of cloud depth below the freezing level and the height of the freezing level. The data used in this study were derived from 70 severe convective storms (34 hail producing and 36 nonhail producing storms) over the Midwestern States. Severe thunderstorms, as used in the development of this technique, were defined as those thunderstorms causing measurable property damage due to strong winds, lightning, or heavy rain. Severe thunderstorms with accompanying hail were defined as those thunderstorms accompanied by hail where hail was listed as the prime cause of property damage even though other phenomena may also have occurred. All tornadoes were excluded from consideration to avoid confusion.

METHOD OF ANALYSIS.–Plot representative upper air soundings (0000Z and 1200Z) on the Skew T diagram. Then analyze the following parameters:

1. Convective condensation level (CCL).

2. Equilibrium level (EL). The EL is found at the top of the positive area on the sounding where the temperature curve and the saturation adiabat through the CCL again intersect. This gives a measure of the extent of the cloud's vertical development and thus an estimate of its top or maximum height.

3. Freezing level. This is defined as the height of the zero degree isotherm.

NOTE: ALL THREE HEIGHTS ARE EXPRESSED IN UNITS OF MILLIBARS.

Next determine the following two parameters:

1. The ratio of the cloud depth below the freezing level (distance in millibars from the CCL to the freezing level) to the cloud's estimated vertical development (distance from the CCL to the EL in millibars) is defined as the cloud depth ratio. For example, if the CCL was at 760 mb, the freezing level at 620 mb, and the EL at 220 mb, then the cloud depth ratio would be computed as follows:

$$\text{Cloud depth ratio} = \frac{140}{540} = .26$$

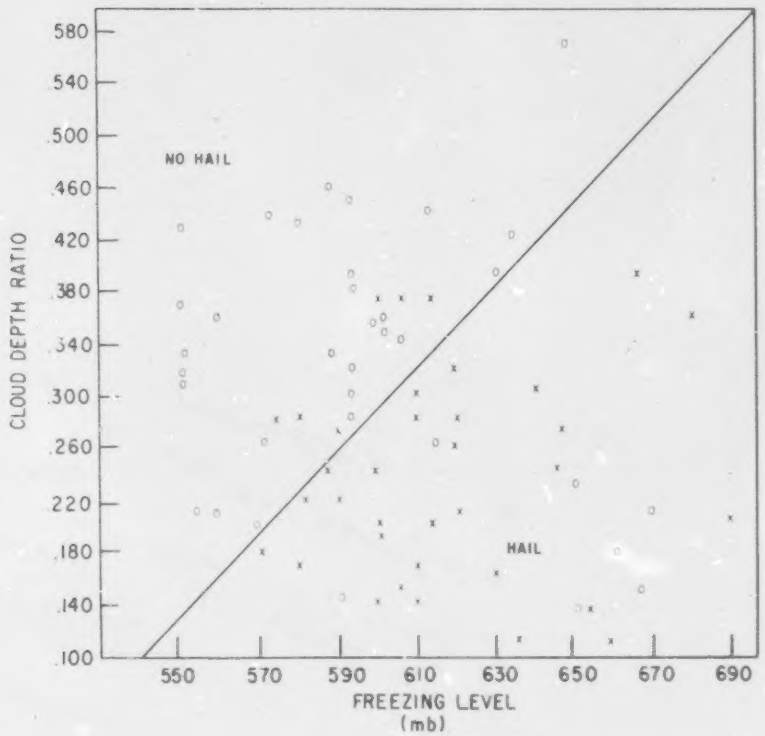

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Figure 11-11.—Scattered diagram showing the distribution of selected hail occurrences at certain midwestern stations during the spring and summer of 1959 and 1960. The freezing level is plotted against the cloud depth ratio. (X = hail reported. 0 = no hail reported.)

2. Height of the freezing level in millibars (620 mb).

With 1 as the ordinate and 2 as the abscissa, enter figure 11-11 for occurrence or nonoccurrence of hail. In this case, hail would be forecast as the value falls in the hail forecast area.

EVALUATION. Using the data in 70 dependent cases, the percent correct for prediction of hail or no hail was 83 percent. This technique combines the two parameters relating hail to convective activity into a single predictor. Although the data used in this study were from the Midwest, the application need not be confined to that area. With some modification of this diagram, this method could serve as a basis for a local forecasting tool for other areas.

Hail Size Forecasting

Once the forecaster has determined that the probability of hail exists, as previously outlined, the next logical question will relate to the size of the hail that may be anticipated. AWS Technical Report 200 (Rev) provides a technique for determining hail size that has been developed by the Air Force. This method utilizes the Skew T Log P diagram.

The first step in forecasting hail is to determine the Convective Condensation Level (CCL). This parameter is evaluated on the adiabatic chart by finding the mean mixing ratio in the moist layer of the lowest 150 mb, and following this saturation mixing ratio line to its intersection with the sounding dry-bulb temperature curve. Next, the moist adiabat through the CCL is traced up to the pressure level where the dry-bulb temperature is −5°C. This pressure level, the dry-bulb temperature curve, and the moist adiabat through the CCL form a triangle outlining a positive area. Figure 11-12 illustrates this procedure. The horizontal coordinate in figure 11-12 is the length of the horizontal side of the triangle in degrees Celsius. The vertical coordinate is the length (in degrees) of a dry adiabat through the triangle. This length is measured from the pressure at the base of the triangle.

These computations for the horizontal length in degrees and altitude in degrees are utilized on the graph in figure 11-13, for the forecast of hailstone diameter.

EXAMPLE OF TECHNIQUE. In the sounding shown in figure 11-12, the CCL is point A. The moist adiabat from the CCL to the pressure level where the temperature is −5°C is the line AB′. The isobar from the point where the air temperature is −5°C to its intersection with the moist adiabat is the line BB′. The dry adiabat from the isobar BB′ through the triangle to the pressure of the CCL is the line HH′. The base of the triangle in degrees Celsius is 6°C (from plus 1 to minus 5). The length of the dry adiabat through the triangle is 21°C (from minus 4 to plus 17). The value on the graph in figure 11-13, with a horizontal coordinate of 6 and a vertical coordinate of 21, is a forecast of 1 inch hail.

In order to more accurately forecast hail size in conjunction with thunderstorms along the Gulf Coast or in any air mass where the Wet-Bulb-Zero height is above 10,500 feet, it is necessary to refer to the graph in figure 11-14. The hail size derived from figure 11-13 is entered on the horizontal coordinate of figure 11-14, and the corrected hail size read off is compatible with the height of the Wet-Bulb-Zero temperature.

A THUNDERSTORM CHECKLIST

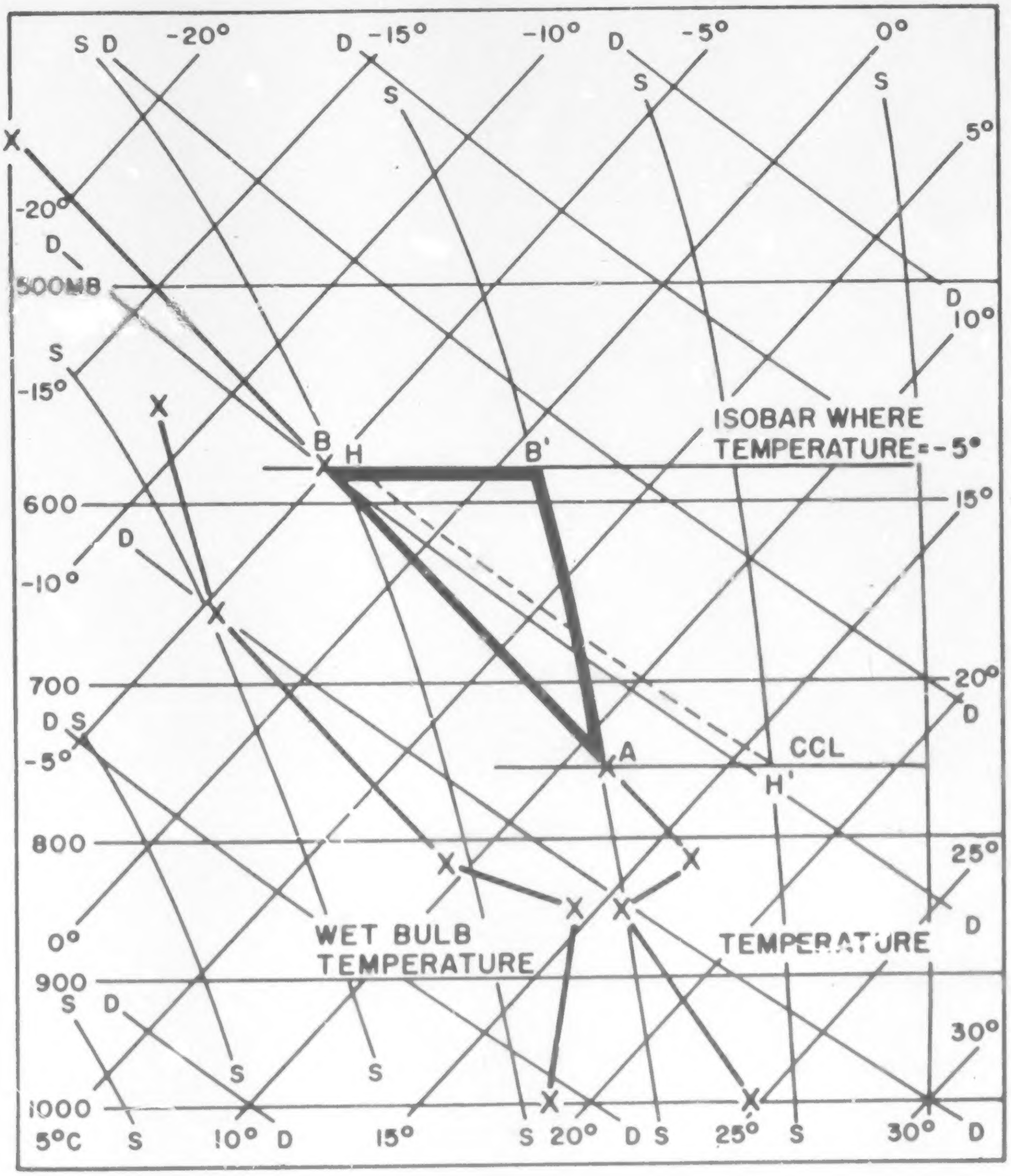
Regardless of where you are forecasting, the factors and parameters favorable for thunderstorm, hail, and gust forecasting should be systematized into some sort of checklist to determine the likelihood and probability of thunderstorms and the attendant weather. Figure 11-15 is a suggested format and checklist to insure that all parameters have been given due consideration.

TORNADOES

Tornadoes are the least understood and least known atmospheric phenomena. Only their relatively small size ranks them as second in the severity of the damage they cause, with tropical cyclones ranking first. Pound for pound, tornadoes are unsurpassed in the damage they cause. Tornadoes are violently rotating columns of air extending downward from a cumulonimbus cloud. They are nearly always observed as funnel clouds. They occur only in certain areas of the world and are most frequent in the United States in the area bounded by the Rockies on the west and the Appalachians on the east. They also occur during certain preferred seasons of the year in the United States with their most frequent occurrence during May. The season of occurrence varies with the locality. Too, 80 percent of the tornadoes in the United States have occurred between noon and 2100.

In the last few years the U.S. Air Force and the National Weather Service have given increasing attention to the forecasting of severe local storms and tornadoes. Since the establishment of the Severe Local Storm (SELS) Forecast Center in Kansas City, Missouri, the predictions of severe thunderstorms and tornadoes have been placed on a more systematic basis. Warnings and forecasts are issued for the United States from this center at periodic intervals for use by both the general public and the military.

Complete details on the forecasting of these phenomena are beyond the scope of this training manual. It is suggested that the reader consult the U.S. Department of Commerce, Forecasting Guide No. 1, Forecasting Tornadoes and Severe Thunderstorms, and the many other excellent texts and publications for a more complete understanding of this problem. The senior Aerographer's Mate should have a basic understanding of the factors leading to the formation of such severe phenomena, to recognize potential situations, and to be able to forecast such phenomena.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X
X
1000
5°C
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10°
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S
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D
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25°
S
30°
D

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Figure 11-12.—Example of hail size forecast sounding.

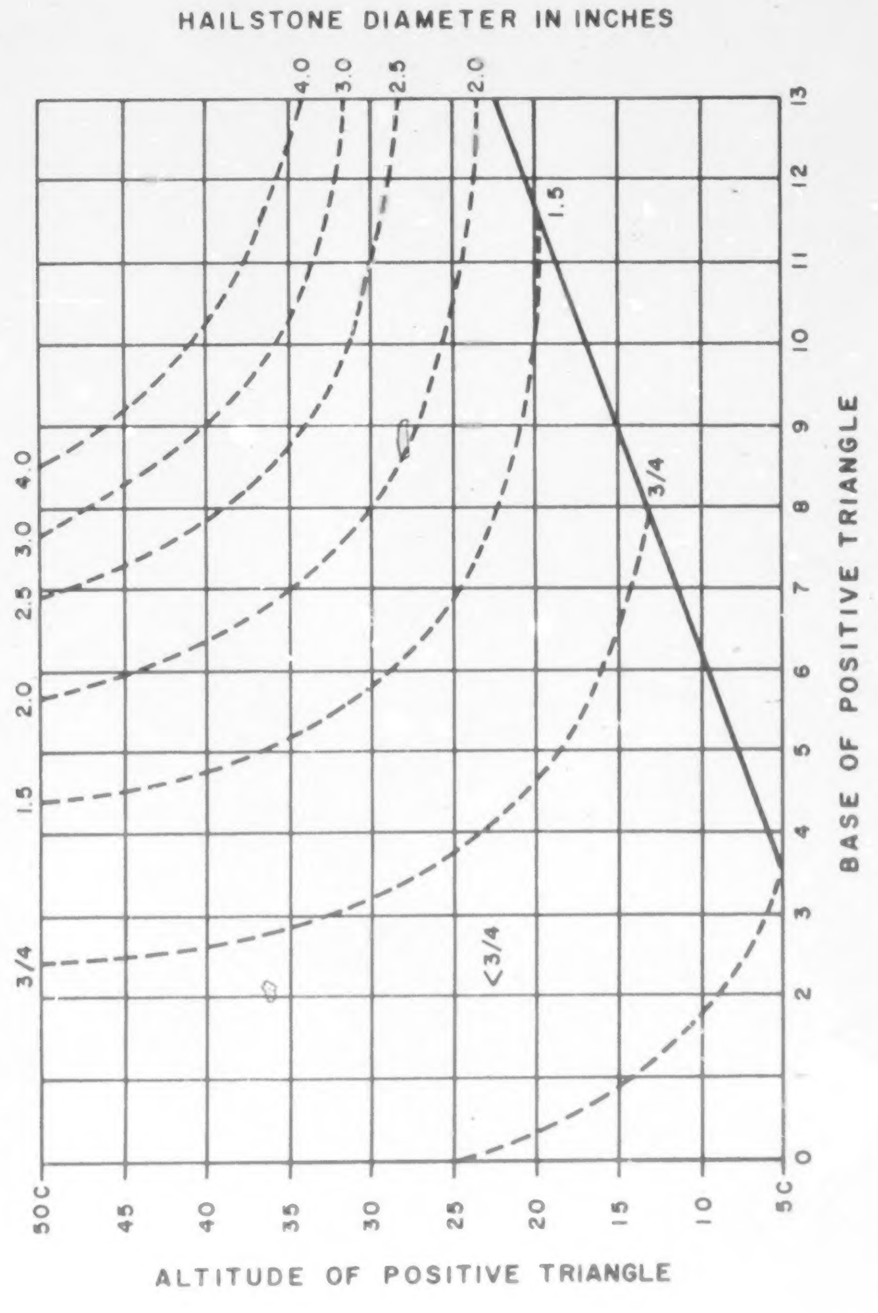

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Figure 11-13.–Fawbush-Miller Hail Graph showing forecast hailstone diameter in inches.

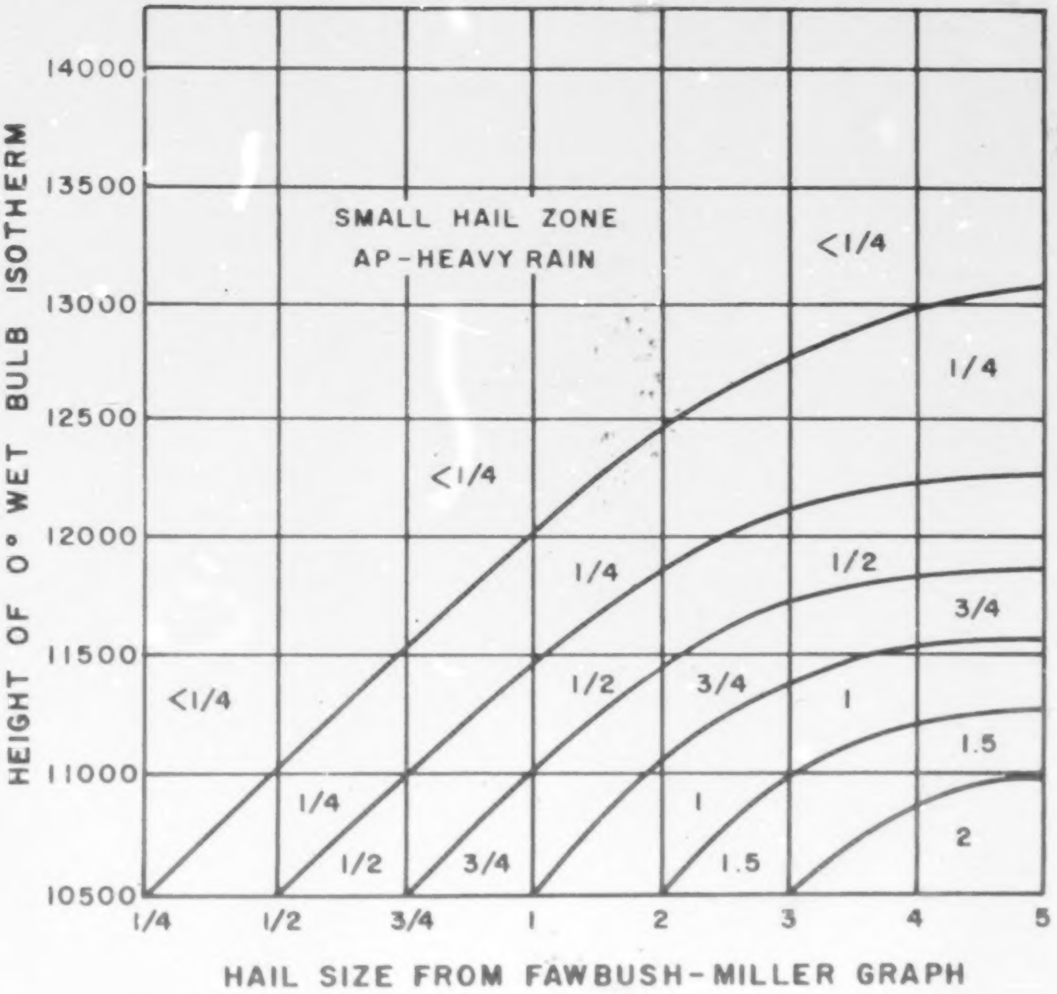

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Figure 11-14.—Hail size at surface expected from tropical air mass thunderstorms.

SURFACE THERMAL PATTERNS AS A FORECASTING AID

This method indicates that thermal tongues averaging 50 miles or less in width are favored locations for tornado activity within the more general area of convective storm activity. These thermal tongues and associated forecasts of tornado activity can be located quite readily on the surface synoptic chart. Use the following technique to supplement the existing tornado forecast.

The procedure to locate areas of potentially severe convective activity with reference to the synoptic surface thermal pattern is as follows:

1. Draw isotherms for every 2 degrees to locate thermal tongues.

WORKSHEET

SUGGESTED CHECK LIST FOR THE DETERMINATION OF AIR MASS THUNDERSTORM ACTIVITY

I. Analyze the Arowagram using the closest sounding & parcel methods

1. Determine the following

a. Convective condensation level (CCL) 5,000'

b. Temperature necessary for convection +28°C(82.4°F)

c. Maximum temperature forecast +30°C (86°F)

d. Is temperature necessary for convection expected to be reached? YES

e. Inversions present? (Stg/Weak/Mdt/Height) NONE

f. Inversions strong enough to prevent or retard convection activity NO

g. Positive energy area favorable/unfavorable FAV

h. Does positive energy area extend well above the freezing level? (Preferably above the -10° C isotherm) YES

i. Does moist layer extend to 10,000 ft? YES

j. At least 3 gr/kg moisture 12,000 ft YES(4.5)

k. Conditionally unstable air extent to 16,000' YES

l. Stability index (fav/unfav) -1 (FAV)

m. Were thunderstorms or clouds of vertical development present previous day? If so, has there been any change at lower or upper levels to retard further development today YES - NO CHANGE

n. Is month and time of year favorable climatologically for thunderstorm development? YES(6.4 PER MO.)

2. Forecast from consideration of a thru n. AFTN TSTMS

Note: The most important single predictor of daytime thunderstorms is moisture in the lower troposphere in the morning.

Determination of f. is of prime consideration. You must decide if heating and lift is sufficient to overcome the inversion.

AG.621

Figure 11-15.—Suggested checklist for the determination of air mass thunderstorm activity.

II. Forecast from using the Bailey Graph SCTD TSTMS

III. Final forecast from all of the above considerations SCTD AFTN

AIR MASS TSTMS BETWEEN 14-1800 LOCAL

IV. Gusts

1. Table 13-1 45 KT
2. AWS Manual Method 40 KT
3. Estimated from reports or expected intensity of storm 38 KT
4. Final Forecast 41 KTS

V. Hail (SURFACE)

1. Yes-No Method YES
2. Fawbush-Miller Method Note if wet bulb zero above 10,500 ft. ½-1"
3. Estimated ½"   4. Forecast ½"

AG.622

Figure 11-15.—Suggested checklist for the determination of air mass thunderstorm activity—Continued.

2. Within the general area in which convective storms are forecast, locate the axis of all pronounced thermal tongues oriented nearly parallel to the gradient flow.

3. On this axis, locate the point with the greatest temperature gradient within 50 to 100 miles to the right, and normal to the flow.

4. From this reference point, a rectangle is constructed with its left side along the axis of the thermal tongue by locating corner points on the axis 25 miles upstream and 125 miles downstream from the reference point. The rectangle is 150 miles long and 50 miles wide.

5. This is the forecast area for possible tornado or funnel cloud development.

TORNADO TYPES

There are commonly believed to be three tornado producing air structures for tornado development over the United States. They are the Great Plains type, the Gulf Coast type, and West Coast type.

Great Plains Type

Tornadoes in this type air mass favor the intersection of the squall line with warm front zones. The tornadoes will generally form on the squall line, hence their prediction involves timely delineation (geographical) of the squall line formation along or in advance of a cold front, upper cold front, or trough. Conditions must favor a downrush of air from aloft. (See chapter 4.)

Gulf Coast Type

In contrast to the air mass type (Great Plains Type) tornadoes also form in an equatorial type air mass that is moist to great heights. Such storms are most common on the coast of the Gulf of Mexico and produce the waterspouts often reported over Florida. Tornadoes are triggered in this air mass primarily by lifting at the intersection of a thunderstorm line with a warm front and less frequently by frontal and prefrontal squall lines.

West Coast Type

Tornadoes also form in relatively cold moist air. This air mass is the Pacific or West Coast type. It is responsible for waterspouts on the West Coast. Tornadoes in this type air mass are normally in a rather extensive cloudy area with scattered rain showers and isolated thunderstorms. Clouds are mostly stratocumulus. Favorable situations for tornado development in this air mass type include the rear of mP cold fronts; well cooled air behind squall lines.

WATERSPOUTS

Phenomena frequently encountered by naval personnel are waterspouts. Waterspouts fall into two classes: tornadoes over water and fair weather waterspouts.

The fair weather waterspout is comparable to a dust devil. It may rotate in either direction, whereas the other type of waterspout rotates cyclonically. In general, waterspouts are not as strong as tornadoes, in spite of the large moisture source and the reduced friction. The water surface under a waterspout is either raised or lowered depending on whether it is affected more by the atmospheric pressure reduction or the wind force. There is less inflow and upflow of air in a waterspout than in a tornado. The waterspout does not lift any significant amount of water from the surface. Ships passing through waterspouts have mostly encountered fresh water.

FORECASTING FOG AND STRATUS

Fog and stratus clouds are hazardous conditions for both aircraft and ship operations. Senior Aerographer's Mates will frequently be called upon to forecast formation, lifting, or dissipation of these phenomena. In order to assist them in providing the best information available we will discuss the various factors that influence the formation and dissipation of fog and stratus, as well as some methods that may be utilized in forecasting these actions.

EFFECT OF AIR MASS STABILITY ON FOG

Fog and stratus are typical phenomena of a warm type air mass. Since a warm type air mass is one with a temperature greater than that of the underlying surface, it is stable, especially in the lower layers.

Through the use of upper air soundings, measurements can be made of temperature and relative humidity, from which stability characteristics can be determined. The publication, Use of the Skew T, Log P in Analysis and Forecasting, NAVAIR 50-IP-5, contains complete information on analyzing upper air soundings and should be consulted by the forecaster.

GENERAL PROCEDURE FOR FORECASTING FOG

One of the basic considerations is that a potential fog or stratus situation exists and that all of the factors are favorable for formation. The particular synoptic situation, time of the year, climatology of the station, stability of the lapse rates, amount of cooling expected, strength of the wind, dewpoint-temperature spread, and trajectory of the air over favorable types of underlying surfaces are all basic considerations to be taken into account.

Consideration of Geography and Climatology

Certain areas are more favorable climatologically for fog formation during certain periods of the year than others. All available information as to the frequency and climatology of fog should be compiled for your station or operating area to determine the times and periods most favorable for formation.

Then determine the location of the station with respect to air drainage or upslope conditions. Next, determine the type of fog to which your location would be exposed. For example, inland stations would be more likely to have radiation fog and shore coastal stations advection fogs. A determination should then be made of air trajectories favorable for fog formation at your station.

Frontal Fogs

Forecasting of these types of fogs is associated with the forecasting of the movement of the fronts and their attendant precipitation areas. For example, fogs can form in advance of a warm front, in the warm air section behind the warm front (when the warm air dewpoint is higher than the cold air temperature), or behind a slow moving cold front when the air becomes saturated.

Air Mass Types

The first step is to determine a trajectory of the air at the station and estimate the changes during the night. If the air has been heated normally during the day and there was no fog the preceding morning, no marked cloud cover during the day, and no trajectory over water, then no fog of marked intensity will form during the night. However, during fall and winter, nights are long and days are short, and conditions are generally stable; and when a fog situation has been in existence, the same conditions tend to remain night after night and the heating during the day is insufficient to effectively raise the temperature of the lower air. Also, determine if the air has had a path over extensive water bodies and whether this path was sufficient to raise the humidity or lower the temperature sufficiently to form fog. Then construct nomograms, tables, etc; using dewpoint depression against time of fog formation for various seasons and winds, and modify these in the light of each particular synoptic situation.

FACTORS TO BE CONSIDERED IN FOG AND STRATUS FORMATION

Wind

Wind velocity is an important consideration in the formation of fog and/or low ceiling clouds.

When the temperature and dewpoint are close at the surface and eddy currents are 100 feet or more in vertical thickness, adiabatic cooling in the upward side of the eddy could give the additional cooling needed to bring about saturation. Any additional cooling would place the air in a temporary supersaturated state. The extra moisture will then condense out of the air, producing a low ceiling cloud. Adiabatic heating on the downward side of the eddy will usually dissolve the cloud particles. If all cloud particles dissolve before reaching the ground, the horizontal visibility should be good. However, if many particles reach the ground before evaporation, the horizontal visibility will be restricted by a moderate fog condition. Clouds forming in eddy current areas may at first be patchy and become identified as scud type clouds. If the cloud forms into a solid layer, it will be a layer of stratus. When conditionally unstable air is present in the eddy area or if the frictional eddy currents are severe enough, stratocumulus clouds will be identified in the area. (See fig. 11-16.)

Saturation of Air

The saturation curve in figure 11-17 shows the amount of moisture in grams per kilogram the air will hold at various temperatures.

The air along the curve is saturated with water vapor and is at its dewpoint. Any further cooling will yield water as a result of condensation, hence fog or low ceiling clouds (depending upon the wind velocity) will form.

Nocturnal Cooling

Nocturnal cooling actually begins after the temperature reaches its maximum during the day. The cooling will continue until sunrise or shortly thereafter. This cooling affects only the lower limits of the atmosphere. If nocturnal cooling reduces the temperature to a value close to the dewpoint, fog, or low ceiling, clouds will

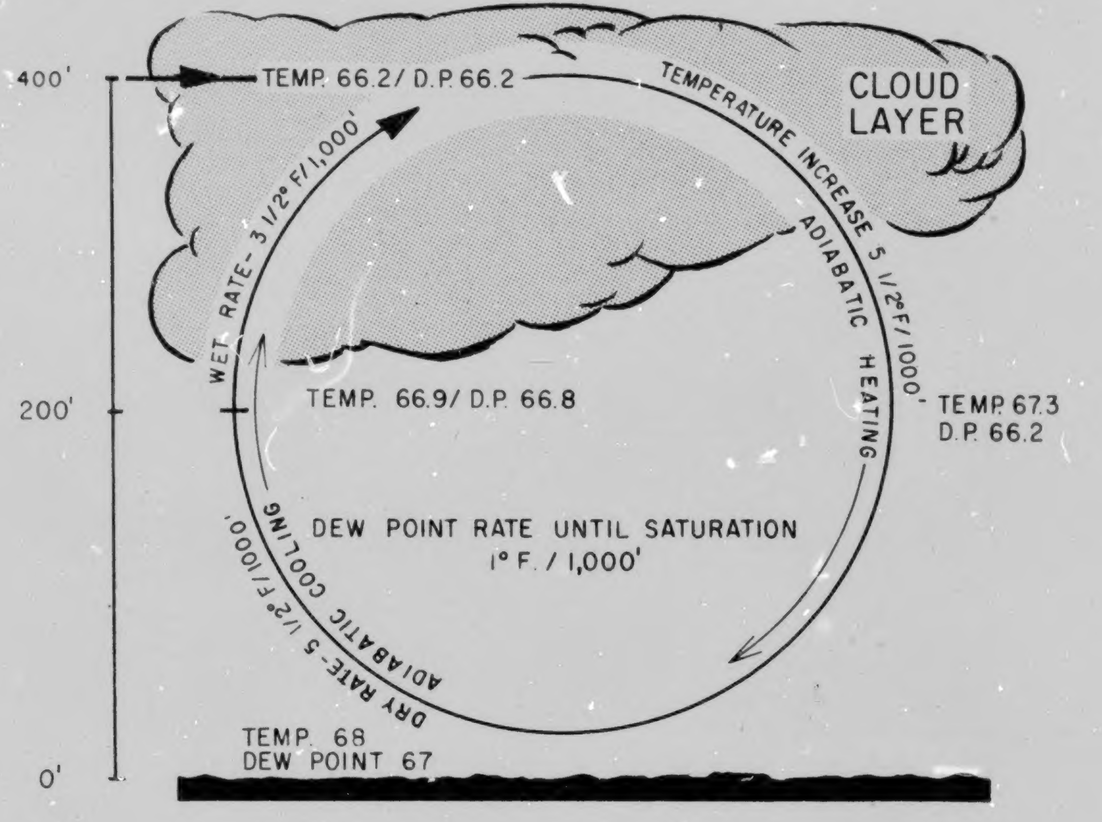
400'
TEMP. 66.2/ D.P. 66.2
TEMPERATURE INCREASE 5 1/2°F/1000'
CLOUD LAYER
WET RATE- 3 1/2° F/1,000'
ADIABATIC HEATING
200'
TEMP. 66.9/ D.P. 66.8
TEMP. 67.3
D.P. 66.2
DEW POINT RATE UNTIL SATURATION
1° F. / 1,000'
DRY RATE- 5 1/2° F/1000'
ADIABATIC COOLING
TEMP. 68
DEW POINT 67
0'

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Figure 11-16.—How wind velocity can cause a low cloud layer.

develop. The wind velocity and terrain roughness will control the depth of the cooled air. Calm wind will allow a patchy type of ground fog or a very shallow continuous ground fog to form. Winds of 5 to 10 knots will usually allow the fog to thicken vertically. Winds greater than 10 knots will usually cause low scud, stratus or stratocumulus to form. (See figs. 11-18 and 11-19 for examples of fog and stratus formation.)

The amount of cooling at night is dependent on soil composition, vegetation, cloud cover, ceiling, and other factors. Cooling may vary over any particular area. A cloud cover based below 10,000 feet has a greenhouse effect on surface temperatures, absorbing some terrestrial radiation and reradiating a portion of this heat energy back to be once again absorbed by the land. This causes a reduction in nocturnal cooling. Nocturnal cooling between 1530 local standard time and sunrise will vary from as little as 5° to 10° with an overcast sky condition based around 1,000 feet to 25° or 30° F with a clear sky or a cloud layer above 10,000 feet. Certain other factors and exceptions must also be considered. If a front is expected to pass the station during the night, or onshore winds are expected to occur during the night, the amount of cooling expected would have to be modified in the light of these developments.

An estimate of the formation time of fog, and possibly stratus, can be aided greatly if some type of saturation time chart such as that illustrated in figure 11-20 can be constructed on

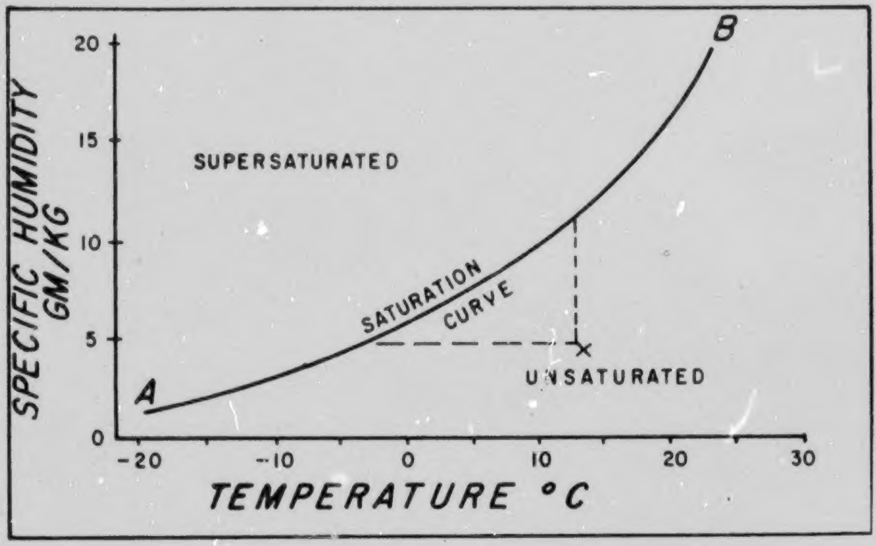

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Figure 11-17.—Saturation curve.

which the temperature forecast versus the dewpoint forecast can be plotted. To use this diagram, note the maximum temperature and consider the general sky condition from the surface map, forecasts, or sequence reports. By projecting the temperature and dewpoint temperature, an estimated time of fog formation can be forecast. If smoke is observed in the area, fog will normally form about 1 hour earlier than the formation line indicates on the charts because of the abundance of condensation nuclei.

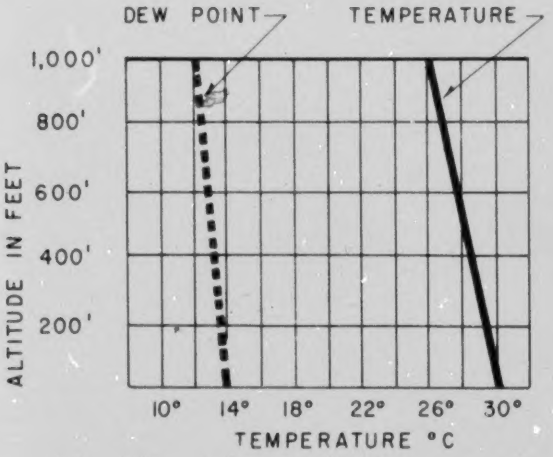
Air Trajectories

The trajectory of the air reaching your station during the forecast period can be another important factor in fog formation.

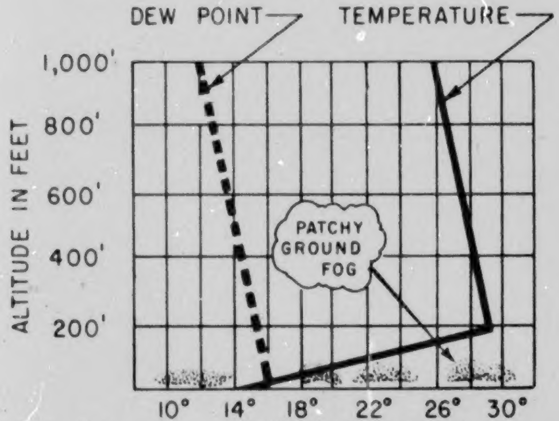
One important factor of air trajectory is warm air moving over a colder surface. This can happen after the rain area of a warm front has moved over the station, and the station is now in the warm sector. Cooling of the air takes place allowing condensation and widespread fog or low stratus to form. To determine the probabilities of condensation behind a warm front, compare the temperature ahead of the front with the dewpoint behind the front. If the temperature ahead of the warm front is lower than the dewpoint behind the front, air behind the front will cool to a temperature near the temperature ahead of the front, causing condensation and the formation of fog or low stratus.

Over water areas, warm air passing over a relatively cold water area may cause enough cooling to allow condensation and the production of low ceiling clouds.

Another instance in which trajectory is important is when cold air moves over a warm water surface, marsh land, or swamp, a shallow layer of fog, called steam fog, may form. In addition, air passing over a wet surface will evaporate a part of the surface moisture, causing an increase in the dewpoint. Whenever there is a moisture source present, air will evaporate a part of this moisture unless the vapor pressure of the air is as great as or greater than the vapor pressure of the water. The dewpoint increase may be enough to allow large eddy currents, nocturnal cooling, or terrain lifting to complete the saturation process and allow condensation to occur.



(A) 1530 LST-MAXIMUM SURFACE TEMPERATURE



(B) SUNRISE-MINIMUM SURFACE TEMPERATURE

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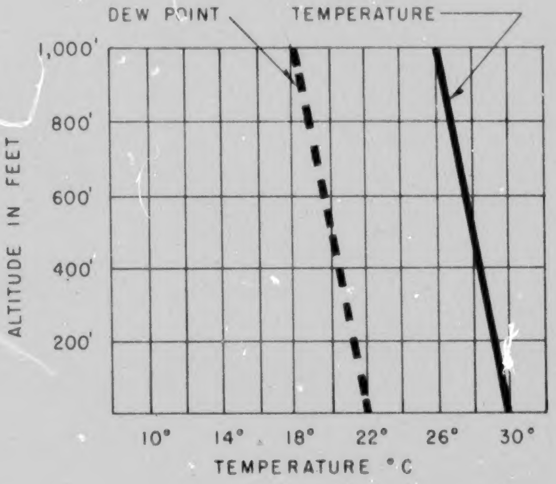
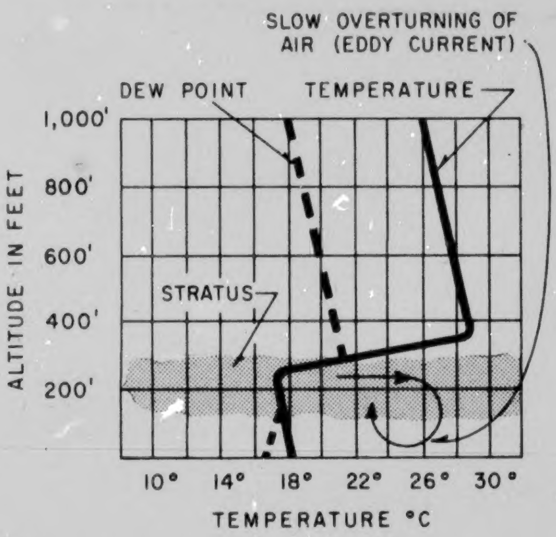
Figure 11-18.—Nocturnal cooling over a land area producing patchy ground fog (calm winds). (A) 1530 LST maximum surface temperature; (B) sunrise-minimum surface temperature.

CONDITIONS FAVORABLE FOR GROUND FOG

In the formation of ground fog, ideally, the air mass would be stable, moist in the lower layers and dry aloft, and under a cloud cover during the day with clear skies at night. Winds should be light, nights long, and the underlying surface wet.



(A) 1530 LST- MAXIMUM SURFACE TEMPERATURE



(B) SUNRISE-MINIMUM SURFACE TEMPERATURE

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Figure 11-19.—Nocturnal cooling over a land area producing stratus (10-15 knot wind). (A) 1530 LST maximum surface temperature; (B) sunrise–Minimum surface temperature.

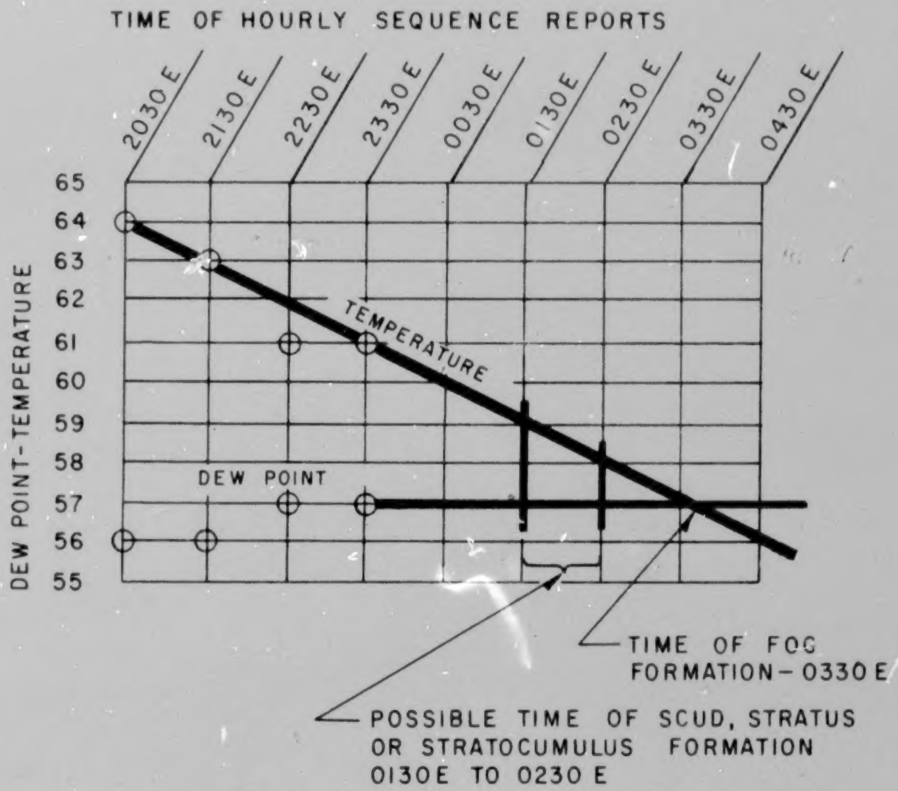

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Figure 11-20.—Saturation time chart. (0's indicate actual temperature and dewpoint observations. Straight line is forecast temperature-dewpoint trends.)

A stationary, subsiding, high-pressure area furnishes the best requirements for light winds, clear skies, stability, and dry air aloft. If the air in the high has been moving over a body of water, or if it lies over ground previously moistened by an active precipitating front, the wet surface will cause an increase in the dewpoint of the lowest layers of the air. In addition, long nights versus short days in fall and winter are favorable for the formation of radiation fog.

CONDITIONS FAVORABLE FOR ADVECTION–RADIATION FOG

Air with a high relative humidity and dewpoint in late summer and early fall around the western side of the Bermuda high or by a return flow from a continental polar high, which has moved out over the water, or by cyclogenesis off the east coast is relatively cold as well as moist. If this air moves inland (replacing warm dry land air), it may be cooled to saturation by nocturnal radiation during the long autumn nights with consequent formation of fog or stratus. The fog is limited to the coastal areas, extending inland between 150 and 250 miles, depending on the wind speed. On the east coast it is limited to the region between the Appalachian Mountains and the Atlantic Ocean.

In late fall and winter, when the continental latitudinal temperature gradient has intensified and the land temperature becomes colder than the adjacent water, the poleward moving air is cooled by advection over colder ground as well as by radiation. If the air is sufficiently moist, fog or stratus may form. During daytime, heating may dissipate the fog or stratus entirely. If not, the heating, together with the wind, which is advecting the air, sets up a turbulence inversion and stratus, or stratocumulus layers form at its base. At night if the air is cooled again and the surface pressure gradient is weak, a surface inversion may replace the turbulence inversion and condensation in the form of fog again occurs at the surface. However, if the pressure gradient is strong, the cooling beneath the inversion intensifies it. Under these conditions stratus or stratocumulus occurs just as in the daytime, except with a lower base.

Late fall and winter advection-radiation fogs can occur any place over the continent that can be reached by maritime air or modified returning continental air. Mainly, this is over the eastern half of the United States. However, since tropical air does not reach as high a latitude in winter as in summer, the frequency of such fog is much smaller in the northern part of the country. With a large, slow moving continental warm high covering the eastern half of the country, however, the fogs may extend all the way from the Gulf of Mexico to Canada.

CONDITIONS FAVORABLE FOR UPSLOPE FOG AND STRATUS

Upslope fog and stratus occur in those regions in which the land slopes gradually upward and which are accessible to invasion by humid stable air masses. In North America the areas best fitting these conditions are the Great Plains of the United States and Canada and the Piedmont east of the Appalachians.

The synoptic conditions necessary for the formation of this type of fog or stratus are the presence of humid air and a wind with an upslope component. The stratus is not advected over the station as a solid sheet. It forms gradually overhead. The length of time between the formation of the first signs of stratus and a ceiling usually ranges from 1/2 hour to 2 hours. Also, under marginal conditions, the stratus may not form a ceiling at all. A useful procedure is to check the hourly observations of surrounding stations, especially those southeastward. If one of these stations starts reporting stratus, the chances of stratus formation at one's own station are high.

CONDITIONS FAVORABLE FOR FRONTAL FOGS

Frontal fogs are of three types: prefrontal (warm front), postfrontal (cold front), and frontal passage.

Prefrontal Fog

Prefrontal (warm-front) fogs occur in stable continental polar (cP) air when precipitating warm air overrides it, the rain raising the dewpoint sufficiently for fog formation. Generally, the wind speeds are slight, and the area most conducive to the formation of this type of fog is one between a nearby secondary low and a low-pressure center. In the entire world the northeastern part of the United States is probably the most prevalent region of this type of fog. They are also of importance along the Gulf and Atlantic coastal plain, in the Midwest, and in the valleys of the Appalachians.

A rule of thumb for forecasting ceiling during prewarm-frontal fog is: If the gradient winds are greater than 25 knots, the ceiling will usually remain 300 feet or higher during the night.

Postfrontal Fog

As with the prefrontal fog, postfrontal (cold-front) fogs are caused by falling precipitation. Fogs of this nature are widespread only when a cold front of an east-west orientation has become quasi-stationary and continental polar air is stable. This type of fog is frequent in the Midwest. Fog or stratiform clouds are prevalent for a considerable distance behind cold fronts associated with stable continental air masses in that region if the cold fronts have produced general precipitation.

Frontal Passage Fog

During the passage of a front, fog may form temporarily if the winds accompanying the front are very light and the two air masses are near saturation. Also, temporary fog may form if the air is suddenly cooled over moist ground with the passage of a precipitating cold front. In low latitudes fog may form in summer if the surface is cooled sufficiently by evaporation of rainwater that fell during frontal passage provided that the moisture addition to the air and the cooling are great enough to cause fog formation.

CONDITIONS FAVORABLE FOR THE FORMATION OF SEA FOGS

Sea fogs are advection fogs which form in warm moist air cooled to saturation as the air moves across cold water. The cold water may occur as a well-defined current or as a gradual latitudinal cooling from south to north. The dewpoint and the temperature of the air undergo a continuous change as the air moves across colder and colder water. The surface air temperature falls steadily, tending to approach the water temperature. The dewpoint also tends to approach the water temperature, but at a slower rate. If the dewpoint of the air is initially higher than the coldest water temperature to be crossed, and if the cooling process continues sufficiently long, the temperature of the air ultimately falls to the dewpoint and fog results. However, if the initial dewpoint is less than the coldest water temperature, the formation of fog is unlikely. Generally in northward moving air, or in air which has previously traversed a warm ocean current, the dewpoint of the air is initially higher than the cold water temperature in the north and fog will form, provided sufficient fetch occurs.

The rate of temperature decrease is largely dependent on the speed at which the air moves across the sea surface isotherms, which in turn is dependent both on the spacing of the isotherms and the velocity of the air normal to them.

The removal of a sea fog from over the coldest water requires a change in air mass (a cold front). A movement of sea fog to a warmer land area leads to rapid dissipation. Upon heating, the fog first lifts, forming a stratus deck; then with further heating, this overcast breaks up into a stratocumulus layer and eventually into convective type clouds, or evaporates entirely. An increase in wind velocity can also raise a surface sea fog into a stratus deck especially if the water is not much, if any, colder than the air. Over very cold water, dense sea fog at the surface may persist even with high winds. Maps are available, showing the mean monthly sea surface isotherms. Synoptic charts show the temperature and dewpoint of the air and its movement. Use of these two charts in conjunction with each other indicates whether the temperature relationship and airflows are such as to make probable the formation or dissipation of fog or stratus.

CONDITIONS FAVORABLE FOR THE FORMATION OF ICE (CRYSTAL) FOGS

When the air temperature is below about -25°F, any water vapor in the air condensing into droplets is quickly converted into ice crystals. A suspension of ice crystals in the air at the surface of the earth is called ice fog. Ice fog occurs mostly in the Arctic regions, and is mainly an artificial fog produced by human activities, occurring locally over settlements and airfields where hydrocarbon fuels are burned. (Burning 1 pound of hydrocarbon fuel produces 1.4 pounds of water.)

When the air temperature is approximately -30°F or lower, ice fog frequently forms very rapidly in the exhaust gases of aircraft, automobiles, or other types of combustion engines. When there is little or no wind, it is possible for an aircraft to generate enough ice fog during landing or takeoff to cover the runway and a portion of the airfield. Depending on the atmospheric conditions, ice fogs may persist for periods which vary from a few minutes to several days.

There is also a fine arctic mist of ice crystals which persists as a haze over wide expanses of the arctic basin during winter; it may extend upward through much of the troposphere—a sort of cirrus cloud reaching down to the ground.

USE OF THE SKEW T DIAGRAM IN FORECASTING THE FORMATION AND DISSIPATION OF FOG

One of the most accepted methods of forecasting the formation and dissipation of fog today makes use of an upper air sounding plotted on the Skew T diagram. The plotting of an upper air sounding is useful in forecasting both the formation and dissipation of fog, but it can be used more objectively in forecasting fog dissipation.

The use of an upper air sounding to determine the possibility of fog formation must be subjective. A study of the existing lapse rate should be made to determine the stability and instability of the lower layers. The surface layer must be stable before fog can form. If it is not found to be stable on the sounding, the cooling expected during the forecast period must be considered, and this modification should be applied to the sounding to determine if the layer will be stable with the additional cooling.

The difference between the temperature and the dewpoint must be considered. If the air temperature and the dewpoint are expected to coincide during the period covered by the forecast, a formation of fog is, of course, very likely.

The expected wind speed must be considered. If the wind speed is expected to be strong, the cooling will not result in a surface inversion favorable for the formation of fog, but may result in an inversion above the surface, which is favorable for the formation of stratus clouds.

DETERMINATION OF FOG HEIGHT

An upper air sounding taken during the time fog is present will show a surface inversion. The fog will not necessarily extend to the top of the inversion. If the temperature and dewpoint have the same value at the top of the inversion, it can be assumed that the fog extends to the top of the inversion. However, if they do not have the same value, the depth of the fog can be determined by averaging the mixing ratio at the surface and the mixing ratio at the top of the inversion. The intersection of this average mixing ratio with the temperature curve is the top of the fog layer.

Two methods that may be used to find the height of the top of the fog layer in feet are reading the height from the pressure-height curve on the Skew T and the dry adiabatic method. In using the pressure-height curve method, locate the point where the temperature curve and the average mixing ratio line intersect on the arowagram. Move this point horizontally until the pressure-height curve is intersected. Determine the height of the fog layer from the value of the pressure-height curve at this level. The dry adiabatic method is based on the fact that the dry adiabatic lapse rate is 1°C per 100 m, or 1°C per 328 ft. Using this method, follow the dry adiabat from the intersection of the average mixing ratio line with the temperature curve to the surface level. Find the temperature difference between the point where the dry adiabat reaches the surface and the point of intersection of the dry adiabat and the average mixing ratio. For example, in figure 11-21 the dry adiabat at the surface is 25° C. The temperature at the intersection of the dry adiabat and the average mixing ratio is 20° C. By applying the dry adiabatic method with a lapse rate of 1° C per 328 ft, we find the height of the top of the fog layer as follows:

$$\frac{1}{328} = \frac{5}{x}$$

$$x = 1{,}640 \text{ ft}$$

Dissipation

To determine the surface temperature necessary for the dissipation of fog, trace dry adiabatically on the Skew T from the intersection of the average mixing ratio line and the temperature curve to the surface level. The temperature of the dry adiabat at the surface level is the temperature necessary for complete dissipation. This temperature is known as the CRITICAL TEMPERATURE. This temperature is an approximation, since it assumes no changes will take place in the sounding from the time of observation to the time of dissipation. This temperature should be modified on the basis of local conditions. (See Fig. 11-21.)

GENERAL DISSIPATION RULES

In considering the dissipation of fog and low ceiling clouds, consideration should be given to the rate at which the ground temperature can increase after sunrise. Vertically thick fog or multiple cloud layers in the area will slow up the morning heating of the ground. It is the heating of the ground that increases the dissipation of the fog overlying the ground. If advection fog is

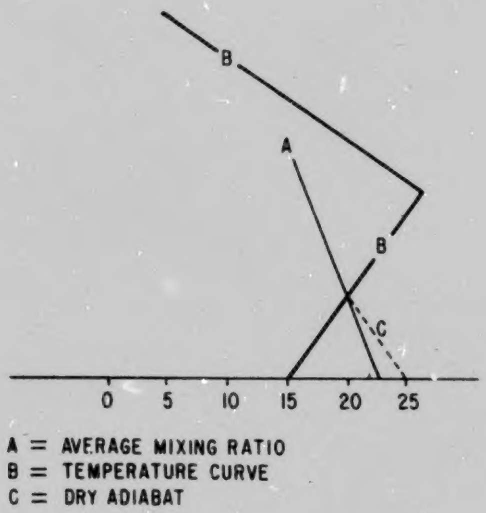

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Figure 11-21.—Dry adiabatic method of determining fog height.

present, the fog may be lifted off the ground to a height where it is classified as stratus. If ground fog is present, the increase in surface air temperature will cause the fog particles to evaporate, thus dissipating the fog. Further heating may evaporate advection fog and low ceiling clouds.

FORECASTING STRATUS FORMATION

The use of the plotted sounding to forecast the formation of stratus can be used in the same manner as it was for forecasting the formation of fog. The lapse rate as it currently exists and modifications which take place must be taken into consideration to determine if a favorable type of inversion will form.

Fog and stratus forecasting are so closely tied together that many of the rules and conditions previously mentioned also apply to stratus.

FORECASTING STRATUS DISSIPATION

A radiosonde sounding taken during the time that stratus exists will show an inversion, usually at a short distance above the surface. The base and the top of this inversion do not necessarily indicate the base and top of the stratus layer.

Determining the Base and Top of a Stratus Layer

One of the first steps in forecasting the dissipation of stratus is to determine the thickness of the stratus layer. The procedure is as follows:

1. Determine a representative mixing ratio between the surface and the base of the inversion.
2. Project this mixing ratio line upward through the sounding.
3. The intersection of the average mixing ratio line with the temperature curve gives the approximate base and maximum top of the stratus layer. Point A in figure 11-22 is the base of the stratus layer, and point B is the maximum top of the layer. Point A is the initial base of the layer; but as heating occurs during the morning, the base will lift. Point B represents the maximum top of the stratus layer, although in the very early morning it might lie closer to the base of the inversion. However, as heating occurs during the day, the top of the stratus layer will also rise and will be approximated by point B. If the temperature and the dewpoint are the same at the top of the inversion, the stratus will extend to this level.

To determine the numerical value of the base and the top of the stratus layer use either the method previously outlined for the fog or the pressure altitude scale.

Determining Dissipation Temperatures

To determine the temperature necessary for the dissipation of the stratus layer to begin and for the dissipation of the stratus layer to be complete, the following steps are followed:

1. From point A in figure 11-22 follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the temperature required to be reached for the stratus dissipation to begin. This is point C.
2. From point B on figure 11-22 follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the surface temperature required for the dissipation

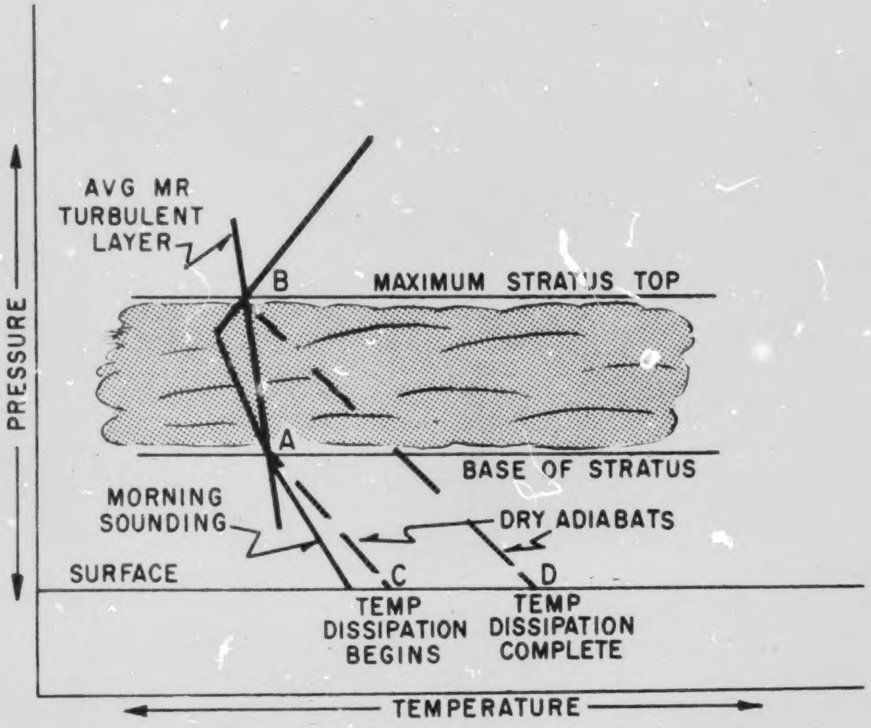
AVG MR
TURBULENT
LAYER
B
MAXIMUM STRATUS TOP
A
BASE OF STRATUS
MORNING
SOUNDING
DRY ADIABATS
SURFACE
C
D
TEMP
DISSIPATION
BEGINS
TEMP
DISSIPATION
COMPLETE
PRESSURE
TEMPERATURE

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Figure 11-22.—Sounding showing the base and the top of stratus layers. Also note temperature at which dissipation begins and temperature when dissipation is complete.

of the stratus layer to be complete. This is point D.

Determining Time of Dissipation

After determination of the temperatures necessary for stratus dissipation to begin and be completed, a forecast of the time these temperatures will be reached must be made. If available, use diurnal heating curves. Otherwise, estimate the length of time for the required amount of heating to take place; and on the basis of this estimate, the time of dissipation may be forecast. Remember to take into consideration the absence or presence of cloud layers above the stratus deck. In addition, the trajectory of the air over the station, if from a water surface, can hold temperatures down for a longer than normal period of time.

One rule of thumb used quite widely in the dissipation of the stratus layer is to estimate the thickness of the layer; and if no significant cloud layers are present above, and normal heating is expected, forecast the dissipation of the layer with an average of 360 feet per hour of heating. In this way an estimate can be made of the number of hours required to dissipate the layer.

FORECASTING ADVECTION FOG OVER THE OCEANS

In the absence of actual temperature and dewpoint data and with a stationary high, the method is as follows (a southerly flow is assumed):

1. Pick out the point on an isobar at which the highest sea temperature is present (either

from the chart of the mean monthly sea temperatures or the synoptic chart), and assume that at this point the air temperature is equal to that of the water and has a dewpoint 2 degrees lower.

2. Find the point on the isobar northward where the water is 2 degrees colder. From this point on, patches of light fog should occur.

3. From a saturation chart such as shown in a previous section of this chapter find how much further cooling would have to occur to give an excess over saturation of 0.4 gr/kg and also 2.0 gr/kg. The first represents the beginning of moderate fog and the second represents drizzle.

4. As the air continues around the northern ridge of the high, it will reach its lowest temperature, and from then on will be subject to warming. The pattern then will be drizzle until the excess is reduced to 2.0 gr/kg and moderate fog until 0.4 gr/kg is reached.

If actual water and temperature data are available, these would be used in preference to climatic mean data. If the high is moving the trajectories of the air particles will have to be calculated.

This method was used with considerable success during World War II to calculate the southern boundary of fog areas. However, the fog is usually less widespread than the calculations indicate and drizzle is less extensive. Also, clearing and lifting on the east side of the high is slightly faster. This method appears to work well in the summer over the Aleutian areas where such fog is frequent.

FORECASTING UPSLOPE FOG

Terrain lifting of the air will cause adiabatic cooling at the dry adiabatic rate of 5 1/2°F per 1,000 feet. If an adequate amount of lifting occurs, fog or low ceiling clouds will form. This is usually a nighttime phenomenon.

The procedures for determining the probabilities of fog or low ceiling clouds during nighttime hours at a station having upslope wind are as follows:

1. Forecast the amount of nocturnal cooling.

2. To determine the expected amount of upslope cooling:

a. Determine the approximate number of hours between sunset and sunrise.

b. Estimate the expected wind velocity for the night hours.

c. Multiply a by b. This will give the distance the upslope wind will move during the period of the day when daylight heating cannot counteract upslope cooling.

d. Determine approximate terrain elevation difference between station and distance computed in c. Elevation difference should be in thousands and tenths of thousands of feet. (Example, 2.5 thousand feet.)

e. Multiply elevation difference by dry adiabatic rate of cooling. (Example, 2.5 times 5.5 = 13.75°F of upslope cooling.)

3. Add expected amount of upslope cooling to expected nocturnal cooling for net expected cooling.

4. Determine 1530 LST temperature-dewpoint spread at station under consideration. If the expected cooling is greater than the 1530 LST spread, either fog or low ceiling clouds should be expected. Wind velocity will determine which of the two conditions will form.

AIRCRAFT ICING

Aircraft icing is another of the weather hazards to aviation. It is important that the pilot be informed about icing because of the serious effects it has on the aircraft's performance. Ice on the airframe decreases lift and increases weight, drag, and stalling speed. In addition, the accumulation of ice on exterior movable surfaces affects the control of the aircraft. If ice begins to form on the blades of the propeller, the propeller's efficiency is decreased and still further power is demanded of the engine to maintain flight. Today, though most aircraft have sufficient reserve power to fly with a heavy load of ice, airframe icing is still a serious problem because it results in greatly increased fuel consumption and decreased range. Further, the possibility always exists that engine-system icing may result in loss of power.

The total effect of aircraft icing is loss of aerodynamic efficiency; loss of engine power; loss of proper operation of control surfaces, brakes, and landing gear; loss of aircrew's outside vision; false flight instrument indications; and loss of radio communication.

For these reasons, it is important that the forecaster be alert for and aware of the conditions conducive for ice formation and be able to accurately forecast the possible formation of icing in flight weather briefings. This chapter covers conditions for the formation of icing, types of icing, intensities of icing, and where icing is most likely to be found as well as basic icing forecasting techniques.

SUPERCOOLED WATER IN RELATION TO ICING

Two basic conditions must be met for ice to form on an airframe in significant amounts. First, the aircraft-surface temperature must be colder than 0°C. Second, supercooled water droplets, liquid water droplets at subfreezing temperatures, must be present. Water droplets in the free air, unlike bulk water, do not freeze at 0°C. Instead, their freezing temperature varies from an upper limit near -10°C to a lower limit near -40°C. The smaller and purer the droplets, the lower is their freezing point. When a supercooled droplet strikes an object such as the surface of an aircraft, the impact destroys the internal stability of the droplet and raises its freezing temperature. In general, therefore, the possibility of icing must be anticipated in any flight through supercooled clouds or liquid precipitation at temperatures below freezing. In addition, frost sometimes forms on an aircraft in clear humid air if both the aircraft and air are at subfreezing temperatures.

PROCESS OF ICE FORMATION ON AIRCRAFT

The first step in ice formation on an aircraft is when the supercooled droplets strike the surface of the aircraft and as the droplet, or portion of it, freezes it liberates the heat of fusion. Some of this liberated heat is taken on by the unfrozen portion of the drop, its temperature is thereby increased, while another portion of the heat is conducted away through the surface in which it lies. The unfrozen drop now begins to evaporate due to its increase in temperature, and in the process it uses up some of the heat which in turn cools the drop. Due to this cooling process by evaporation the remainder of the drop is frozen. Icing at 0°C will occur only if the air is not saturated. This is due to the fact that the nonsaturated condition is favorable for evaporation of part of the drop. Evaporation of the drop cools the drop below freezing, and ice formation then can take place.

TYPES OF ICING

There are three basic forms of ice accumulation on aircraft: rime ice, clear ice, and frost. In addition, mixtures of rime and clear ice are common. The type of icing, clear or rime, is dependent primarily upon the droplet size.

Some of the factors which determine icing types are: droplet size, liquid water content (the amount of water in the form of droplets in a given volume of air), temperature, air speed, and size and shape of the airfoil.

Among those factors favorable for dense structure (like clear ice) are: High water content, large droplet size, temperature only slightly below freezing, high air speed, and thin airfoil. Because of these numerous variables, it should be stressed that many gradations in appearance, type, and structure of icing may often be found on the same aircraft.

Rime Ice

Rime ice is a rough milky, opaque ice formed by the instantaneous freezing of small supercooled droplets as they strike the aircraft. The fact that the droplets maintain their nearly spherical shape upon freezing and thus trap air between them gives the ice its opaque appearance and makes it porous. Rime ice occurs at temperatures below -8° to -10°C, although it has been observed between -2° and -28°C. Rime ice is most frequently encountered in stratiform clouds. In comparison to clear ice, it is relatively easy to get rid of by conventional methods, even though it distorts the airfoil to a larger degree than does clear ice.

Clear Ice

Clear ice is a glossy, clear, or translucent ice formed by relatively slow freezing of large supercooled droplets. The large droplets spread out over the airfoil prior to complete freezing, forming a sheet of clear ice. As a result of the spreading of this supercooled water and its slow freezing, very few air bubbles are trapped within the ice, which accounts for its clearness. Although clear ice is expected mostly with temperatures between 0° and -10°C, it does occur with temperatures as cold as -25°C. The atmospheric conditions which produce clear ice are encountered most frequently in cumuliform clouds. Clear ice also forms rapidly on aircraft while flying in zones of freezing rain or drizzle.

Frost

Frost is a deposit of a thin layer of crystalline ice that forms on the exposed surfaces of parked aircraft when the temperature of the exposed surface is below freezing (while the temperature of the free air may be above freezing). The deposit of ice forms during the night radiational cooling, in a manner similar to the formation of hoar frost on the ground. Frost may also form on aircraft in flight when a cold aircraft descends from a zone of freezing temperatures into a warmer moist layer below. Frost can cover the windshield or canopy and completely restrict outside vision. During penetration for landing, when visible moisture or high relative humidity exists, frost may form on the inner side of the canopy. This can be a definite hazard if no preventive action is taken.

Frost is a deceptive form of icing. It affects the lift-drag ratio of an aircraft, and is a definite hazard during takeoff. All frost should be removed from the aircraft prior to taking off.

INTENSITIES OF ICING

There are four intensities of aircraft icing that are defined meteorologically. The definition of each was standardized by Naval Weather Service Command Instruction 3140.4.

Trace

Ice becomes perceptible. Rate of accumulation slightly greater than the rate of sublimation. It is not hazardous even though deicing/anti-icing equipment is not used, unless encountered for an extended period of time (over 1 hour).

Light

The rate of accumulation may create a problem if flight is prolonged in this environment (over 1 hour). Occasional use of deicing/anti-icing equipment removes/prevents accumulation. It does not present a problem if the deicing/anti-icing equipment is used.

Moderate

The rate of accumulation is such that even short encounters become potentially hazardous and use of deicing/anti-icing equipment or diversion is necessary.

Severe

The rate of accumulation is such that deicing/anti-icing fails to reduce or control the hazard. Immediate diversion is necessary.

ICING HAZARDS NEAR THE GROUND

Certain icing hazards exist on or near the ground. One hazard results when wet snow is falling during takeoff. This situation can exist when free air temperature at the ground is near 0°C. The wet snow sticks tenaciously to aircraft components, and freezes when the aircraft encounters markedly colder temperatures during the climb.

If not removed before takeoff, frost, sleet, frozen rain, and snow accumulated on parked aircraft are operational hazards. Another hazard arises from the presence of puddles of water, slush, and/or mud on airfields. When the air temperature on the airframe is colder than 0°C, water blown by the propellers or splashed by wheels can form ice on control surfaces and windows. Freezing mud is particularly dangerous because the dirt may clog controls and cloud the windshield.

PHYSICAL FACTORS AFFECTING AIRCRAFT ICING

The amount and rate of icing depend on a number of meteorological and aerodynamic factors, including temperature, the amount of liquid water in the path of the aircraft, the fraction of this liquid water collected by the aircraft (the collection efficiency), and the amount of aerodynamic heating.

Temperature

Temperature has a direct effect on the fraction of water which freezes instantaneously on impact. When a large droplet strikes the aircraft with temperature just below freezing, a fraction of the water is carried off the surface and a portion freezes. In the case of smaller droplets, or large droplets at a slightly lower temperature, the entire droplet may freeze on the aircraft. The temperature range in which ice is most likely to form is generally from 0° to -20°C with the lower limit in the vicinity of -40°C. Supercooled water droplets have been found at temperatures colder than -40°C and at altitudes above 40,000 feet.

Airflow around aircraft surfaces can decrease the free air temperature in some cases as much as 2°C. This explains the fact that icing sometimes takes place when the outside free air temperature gage indicates temperature slightly above freezing.

Liquid-Water Content

Under icing conditions, the liquid-water content in the cloud is the most important parameter in determining the ice accumulation rate. In general, the lower and warmer the base of the cloud, the higher is its water content. Within the cloud, the average liquid-water content increases with altitude to a maximum value and then decreases. The maximum concentration usually occurs at a lower level in stratiform than in cumuliform clouds, and the average liquid-water content is usually less than that of a cumuliform cloud.

Droplet Size

Droplet size affects the collection efficiency and hence the icing rate. The size distribution and median size of the droplets in a cloud are related to type, depth, and age of the cloud; to the strength of the updrafts; to the humidity of the air mass; and to other factors. Since both the liquid-water content and droplet size are generally greater in cumuliform clouds than other cloud types, it would seem logical that cumuliform clouds should be particularly conducive to icing. However, the effect exerted by other variables, such as the speed, shape, and size of the aircraft components, may be sufficiently great for meteorological conditions leading to light icing for one type aircraft, to result in moderate or severe icing for another.

Collection Efficiency

The icing rate depends to a large degree upon the collection efficiency of the aircraft component involved. The fraction of liquid water collected by the aircraft varies directly with droplet size and aircraft speed, and inversely with the size or geometry of the collecting surface. Those components having a large radius (canopies, thick wings, etc.) deform the airflow more, permitting only a small portion of the droplets to be caught on the surface.

Aerodynamic Heating

This is a rise in temperature resulting from adiabatic compression and friction as the aircraft penetrates the air. The amount of heating varies primarily with the speed of the aircraft and the altitude (air density), and may range from approximately 1 degree for very slow aircraft at low altitudes to more than 50 degrees for supersonic jets at low altitudes. Thus, although it is frequently stated that an aircraft flying through any supercooled liquid-water cloud must anticipate icing, it actually is necessary that the amount of supercooling exceed the amount of aerodynamic heating.

DISTRIBUTION OF ICING IN THE ATMOSPHERE

The atmospheric distribution of potential icing zones is mainly a function of temperature and cloud structure. These factors, in turn, vary with altitude, synoptic situation, orography, location, and season.

Altitude and Temperature

Aircraft icing is generally limited to the layer of the atmosphere lying between the freezing level and the -40°C isotherm. However, icing has occasionally been reported at temperatures colder than -40°C is the upper parts of cumulonimbus and other clouds. In general, the frequency of icing decreases rapidly with decreasing temperature, becoming rather rare at temperatures below -30°C. The normal vertical temperature distribution is such that icing is usually restricted to the lower 30,000 feet of the troposphere. The types of icing associated with temperature ranges will generally be: clear, 0°C to -10°C; a mixture of clear and rime -10°C to -15°C; rime -15°C to -20°C, with possible rime at lower temperatures.

Icing in Relation to Cloud Type

Stable air masses often produce stratus type clouds with extensive areas of relatively continuous icing conditions. Unstable air masses generally produce cumulus clouds with a limited horizontal extent of icing conditions, where the pilot can expect the icing conditions, where the pilot can expect the icing to become more severe at higher altitudes in the clouds. The type and amount of icing will vary considerably with each type cloud.

STRATIFORM CLOUDS.—Icing in middle- and low-level stratiform clouds is confined, on the average, to a layer between 3,000 and 4,000 feet thick. However, pilots frequently encounter situations where multiple layers of clouds are so close together that visual navigation between layers is not feasible. In such cases the maximum depth of continuous icing conditions rarely exceeds 6,000 feet. The intensity of icing generally ranges from light to moderate with the maximum values occurring in the upper portions of the cloud. Both rime and mixed icing are observed in stratiform clouds. The main hazard lies in the great horizontal extent of some of these cloud decks. High-level stratiform clouds are composed mostly of ice crystals and give little icing.

CUMULIFORM CLOUDS.—The zone of probable icing in cumuliform clouds is smaller horizontally but greater vertically than in stratiform clouds. Icing is more variable in cumuliform clouds because many of the factors conducive to icing depend to a large degree on the stage of development of the particular clouds. Icing intensities may range from generally light in small supercooled cumulus to moderate or severe in cumulus congestus and cumulonimbus. The most severe icing occurs in cumulus congestus clouds just prior to their change to cumulonimbus. Although icing occurs at all levels above the freezing level in a building cumulus, it is most intense in the upper half of the cloud. Icing is generally restricted to the updraft regions in a mature cumulonimbus, and to a shallow layer near the freezing level in a dissipating thunderstorm. Icing in these type clouds is usually clear or mixed.

CIRRIFORM CLOUDS.—Aircraft rarely occurs in cirrus clouds, even though some of them do contain a small proportion of water droplets. However, icing of moderate intensity has been reported in the dense cirrus and anvil-tops of cumulonimbus, where updrafts may maintain considerable water at rather low temperatures.

Icing in Frontal Zones

About 85 percent of all icing conditions reported are associated with frontal zones. Aerographer's Mates, therefore, should have a thorough knowledge of the danger areas in the frontal zones so that pilots may be properly briefed. The basic approach to forecasting icing in frontal zones is to relate the fronts to the cloud types in them.

Along warm fronts the stratified cloud system associated with the upglide of warm, moist air over a warm front often reaches dangerous subzero temperatures, where rime or glaze icing will occur. If the warm air mass is conditionally unstable, cumuliform clouds may form. These clouds are favorable to the formation of glaze of rime, or a combination of them. Figure 11-23 shows a typical warmfront structure, with the most probable icing zones and a possible flightpath to minimize icing. The cloud system of a warm front is extensive. A flight through such a system is a long one and the icing dangers are therefore increased.

Cumuliform clouds are associated with cold fronts. Glaze and mixed icing must therefore be expected in them. The cloud systems are usually relatively narrow as compared with those of a warm front, and the time spent flying through

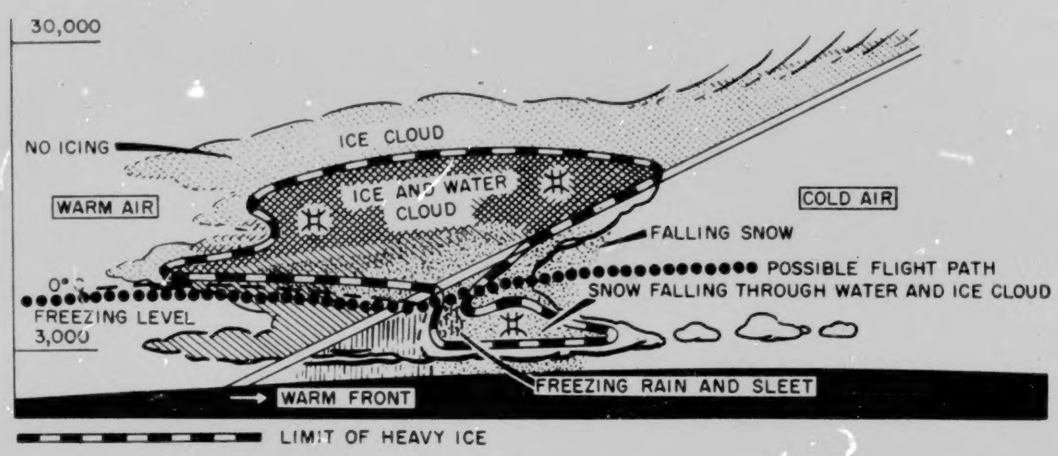
30,000
NO ICING
ICE CLOUD
ICE AND WATER CLOUD
WARM AIR
COLD AIR
FALLING SNOW
POSSIBLE FLIGHT PATH
SNOW FALLING THROUGH WATER AND ICE CLOUD
0°C
FREEZING LEVEL
3,000
FREEZING RAIN AND SLEET
WARM FRONT
LIMIT OF HEAVY ICE

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Figure 11-23.—Warm-front icing condition.

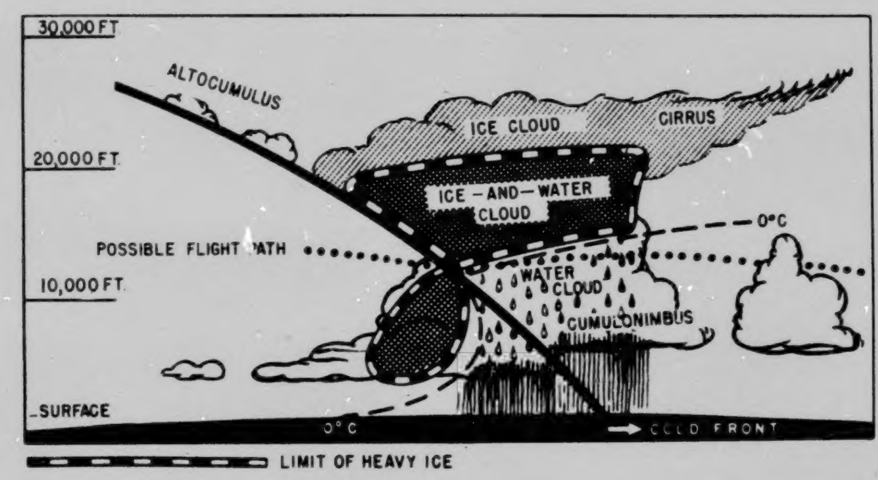
30,000 FT
ALTOCUMULUS
ICE CLOUD
CIRRUS
20,000 FT
ICE-AND-WATER CLOUD
0°C
POSSIBLE FLIGHT PATH
10,000 FT
WATER CLOUD
CUMULONIMBUS
SURFACE
0°C
COLD FRONT
LIMIT OF HEAVY ICE

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Figure 11-24.—Icing zones along a cold front.

them is shorter. However, due to the heavy precipitation, icing can be rapid and extremely dangerous, even though it is of short duration. Figure 11-24 illustrates a typical cold-front icing situation.

Occluded fronts present less of an icing hazard because precipitation has been occurring for some time, and the cloud system contains less water. However, since the cloud system is extensive, accumulation can still be dangerous.

Orographic Influences

The lifting of conditionally unstable air over mountain ranges is one of the most serious ice-producing processes experienced in the United States. When mT air moves northward and eastward over the Appalachian Mountains, it is often cooled to subfreezing temperatures. An icing hazard exists for all air traffic that must travel through this area. Similarly, mP air in winter, approaching the west coast of the United States, contains considerable moisture in its lower levels. As it is forced aloft by the successive mountain ranges encountered in its eastward movement, severe icing zones develop.

The movement of a front across a mountain brings together two important factors that aid in the formation of icing zones. A study of icing in the Western United States has shown that almost all the ice cases occurred where the air was blowing over a mountain slope or up a frontal surface or both.

The most severe icing zones over mountaintops will be above the crests and slightly to the windward side of the ridge. Usually the icing zones extend about 4,000 feet above the tops of the mountains. In the case of unstable air, they may extend higher.

OPERATIONAL ASPECTS OF AIRCRAFT ICING

Due to the large number of types and different configurations of aircraft, this discussion is limited to the general types of aircraft rather than specific models.

Reciprocating-Engine Aircraft

Because of their relatively low speed and service ceilings, this type of aircraft is more susceptible to icing for long periods of time than jet aircraft. Due to the thick wings, canopies, and other features, the collection efficiency is smaller than those of the trimmer and faster turbojet aircraft. However, the hazard is greater because the slower speeds produce less aerodynamic heating and because they operate for longer periods of time in altitudes and areas more conducive to icing.

Propeller icing is a very dangerous form of icing for this type aircraft because of the tremendous loss of power and vibration. Propeller icing varies along the blade due to the differential velocity of the blade causing a temperature increase from the hub to the propeller tip. Modern propellers have deicers of various kinds on them. However, these deicers are curative, not preventive, and the danger remains.

Another serious icing condition exists near the airscoops, carburetor inlet screens, and other induction systems.

Carburetor icing is treacherous. It occurs under a wide range of temperatures and can result in complete engine failure. Carburetor ice forms during vaporization of fuel combined with the expansion of air as it passes through the carburetor. Temperature drop in the carburetor can be as much as 40°C, but is usually 20°C or less. The temperature at which carburetor icing will form depends upon many factors such as relative humidity, type of gas and its ingredients, and the type of carburetors.

Turbojet Aircraft

These high speed aircraft generally cruise at altitudes well above levels where severe icing exists. The greatest problem will be on takeoff, climb, approach, or go-around-because of the greater probability of encountering supercooled water droplets at low altitudes. Also, the reduced speeds result in a decrease of aerodynamic heating.

Turbojet engines experience icing both externally and internally. All exposed surfaces are subject to airframe ice as well as is the inlet ducting and internal elements. Icing can occur in the inlet duct at temperatures well above freezing when the aircraft is operating at slow, low altitude speeds or on the ground.

The jet aircraft usually has little concern about structural icing except possibly on landings, takeoffs, climbs and when operating at slow speeds at low altitudes.

Internal icing poses special problems to jet aircraft and jet engines. In flights through clouds which contain supercooled water droplets, air intake duct icing is similar to wing icing. However, the ducts may ice when skies are clear and temperatures are above freezing. While taxiing, and during takeoff and climb, reduced pressure exists in the intake system which lowers temperatures to a point that condensation and/or sublimation takes place, resulting in ice formation. This temperature change varies considerably with different types of engines. Therefore, if the free air temperature is 10°C or less (especially near the freezing point) and the relative humidity is high, the possibility of induction icing definitely exists.

Turboprop Aircraft

The problems of aircraft icing for this type of aircraft combine those associated with conventional aircraft and turbojet aircraft. Engine icing problems are similar to those encountered by turbojet aircraft while propeller icing is similar to that encountered by conventional aircraft.

Rotary-Wing Aircraft

When helicopters do encounter icing conditions, which would either be in IFR conditions or when flying through an area of freezing rain or drizzle, they are similar but more hazardous than fixed wing aircraft. When icing forms on the rotor blades while hovering, conditions become hazardous because the helicopter is operating near peak operational limits. Icing also affects the tail rotor, control rods and links, and air intakes and filters.

PRELIMINARY CONSIDERATIONS

The first phase of the procedure in the preparation of an aircraft icing forecast consists of making certain preliminary determinations. These are essential regardless of the technique employed in making the forecast.

Clouds

Determine the present and forecast future distribution, type, and vertical extent of clouds along the flightpath. Clouds can be analyzed and forecast using the information contained in surface weather observations, radiosonde observations, pilot reports, and surface and upper air charts using synoptic models, physical reasoning, and empirical studies. The influences of local effects such as terrain features and others should not be overlooked.

Temperatures

Determine those segments of the proposed flightpath which will be in clouds colder than 0°C. A reasonable estimate of the freezing level can be made from the data contained on freezing level charts, constant-pressure charts, radiosonde, reconnaissance, and AIREP observations, or by extrapolation from surface temperatures.

Precipitation

Check surface reports and synoptic charts for precipitation along the proposed flightpath, and forecast the precipitation character and pattern during the flight—special consideration should be given to the possibility of freezing precipitation.

Note that each of the following methods and forecast rules assumes that two basic conditions must exist for the formation of icing. These are that the surface of the aircraft must be colder than 0°C, and that supercooled liquid-water droplets, clouds, or precipitation must be present along the flightpath.

ICING FORECAST

Icing Intensity Forecasts From Upper Air Data

Check upper air charts, pilot or reconnaissance reports, and radiosonde reports for the dewpoint spread at the flight level. Also check the upper air charts for the type of temperature advection along the route. In one study considering only the dewpoint spread aloft, it was found that there was an 84 percent probability that there would be no icing if the spread were greater than 3°C, and an 80 percent probability that there would be icing if the spread were less than 3°C.

The type of thermal advection or the presence of building cumuliform clouds taken in conjunction with the dewpoint spread showed a definite association between the two. When the dewpoint spread was 3°C or less in areas of warm air advection at flight level, there was a 67 percent probability of no icing and a 33 percent probability of light or moderate icing. However, with a dewpoint spread of 3°C or less in a cold frontal zone, the probability of icing reached 100 percent. There was also a 100 percent probability of icing in building cumuliform clouds when the dewpoint spread was 3°C or less. With the spread greater than 3°C, light icing was probable in about 40 percent of the region of cold air advection with a 100 percent probability of no icing in regions of warm or neutral advection. However, it appears on the basis of further experience that a more realistic spread of 4°C at temperatures near -10° to -15°C should be indicative of probable clouds, and that spreads of about 2° or 3°C should be indicative of probable icing. At other temperatures use the values in the following rules:

1. If the temperature is:
   a. 0° to -7°C, and the dewpoint spread is greater than 2°C, there is an 80 percent probability of no icing under these conditions.
   b. -8° to -15°C, and the dewpoint spread is greater than 3°C, forecast no icing with an 80 percent chance of success.
   c. -16° to -22°C, and the dewpoint spread is greater than 1°C, a forecast of no icing would have a 90 percent chance of success.
   d. Colder than -22°C, forecast no icing regardless of what the dewpoint spread is with 90 percent probability of success.
2. If the dewpoint spread is 2°C or less at temperature of 0° to -7°C, or is 3°C or less at -8° to -15°C:
   a. In zones of neutral or weak cold air advection, forecasting light icing (75 percent probability).
   b. In zones of strong cold air advection, forecast moderate icing (80 percent probability).
   c. In areas with vigorous cumulus buildups due to insolational surface heating, forecast moderate icing.

Icing Intensity Forecasts From Surface Chart Data

If upper air data and charts are not available, check the surface charts for locations of the cloud shields of fronts, low-pressure centers, and precipitation areas along the route.

Icing Intensity Forecasts From Precipitation Data

Within clouds not resulting from frontal activity or orographic lifting, over areas with steady nonfreezing precipitation, forecast little or no icing. Over areas without steady nonfreezing precipitation, particularly cumuliform clouds, forecast moderate icing.

Icing Intensity from Clouds Due To Frontal or Orographic Lifting

Within clouds resulting from frontal or orographic lifting, neither the presence nor the absence of precipitation can be used as indicators of icing.

1. Within clouds up to 300 miles ahead of the warm front surface position, forecast moderate icing.
2. Within clouds up to 100 miles behind the cold front surface position, forecast severe icing.
3. Within clouds over a deep, almost vertical low-pressure center, forecast severe icing.
4. In freezing drizzle, below or in clouds, forecast severe icing.
5. In freezing rain below or in clouds, forecast severe icing.

Types of Icing Forecast

The foregoing rules refer only to the intensity of icing and not the type. The following rules apply to the type of icing:

1. Forecast rime icing when the temperatures at flight altitudes are colder than -15°C, or when between -1° and -15°C in stable stratiform clouds.
2. Forecast clear icing when temperatures are between 0° and -8°C in cumuliform clouds and freezing precipitation.
3. Forecast mixed rime and clear icing when temperatures are between -9° and -15°C in unstable clouds.

Example of Forecast of Icing From Upper Air Diagram

It has been previously pointed out how the thickness of clouds, as well as the top of the overcast, may be estimated with accuracy from the dewpoint depression, its behavior, and from changes in the lapse rate.

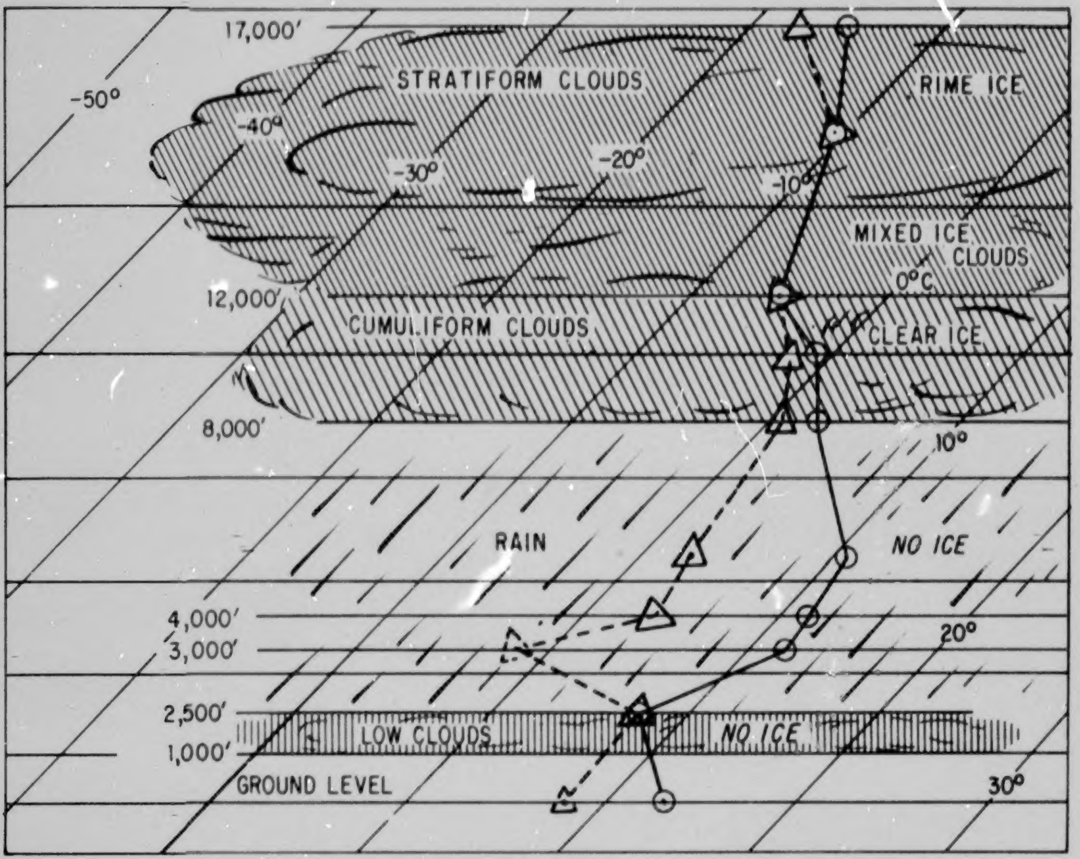
17,000′
-50°
STRATIFORM CLOUDS
RIME ICE
-40°
-30°
-20°
-10°
MIXED ICE
CLOUDS
0°C
12,000′
CUMULIFORM CLOUDS
CLEAR ICE
8,000′
10°
RAIN
NO ICE
4,000′
3,000′
20°
2,500′
LOW CLOUDS
NO ICE
1,000′
GROUND LEVEL
30°

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Figure 11-25.—An illustration of the analysis of cloud type and icing type from an upper air diagram.

The analysis of cloud type and icing type from an upper air diragram is illustrated in figure 11-25. First look at the general shape of the curve. The most prominent feature is the inversion, showing a very stable layer between 2,500 and 3,000 feet. Note that the dewpoint depression is less than one degree at the base of the inversion. Moisture exists in a visible form at this dewpoint depression, so expect a layer of broken or overcast clouds whose base will be approximately 2,000 feet and will be topped by the base of the inversion. The next most prominent feature is the high humidity, as reflected by the dewpoint depressions, at 8,000 feet. Since the dewpoint depression is 2°C, a probability of clouds exists at this level. There is a rapid increase in the dewpoint depression at 17,000 feet indicating that the top of the cloud layer is at this level. We would then assume the cloud layer existed between 8,000 and 17,000 feet. The next step is to determine, if possible, the type of clouds in between these levels. If this sounding were plotted on the Skew T, you would be able to see that the slope of the lapse rate between 8,000 and 12,000 feet is shown to be unstable by comparison to the nearest moist adiabat. The clouds will then display unstable cumuliform characteristics. Above 12,000 feet

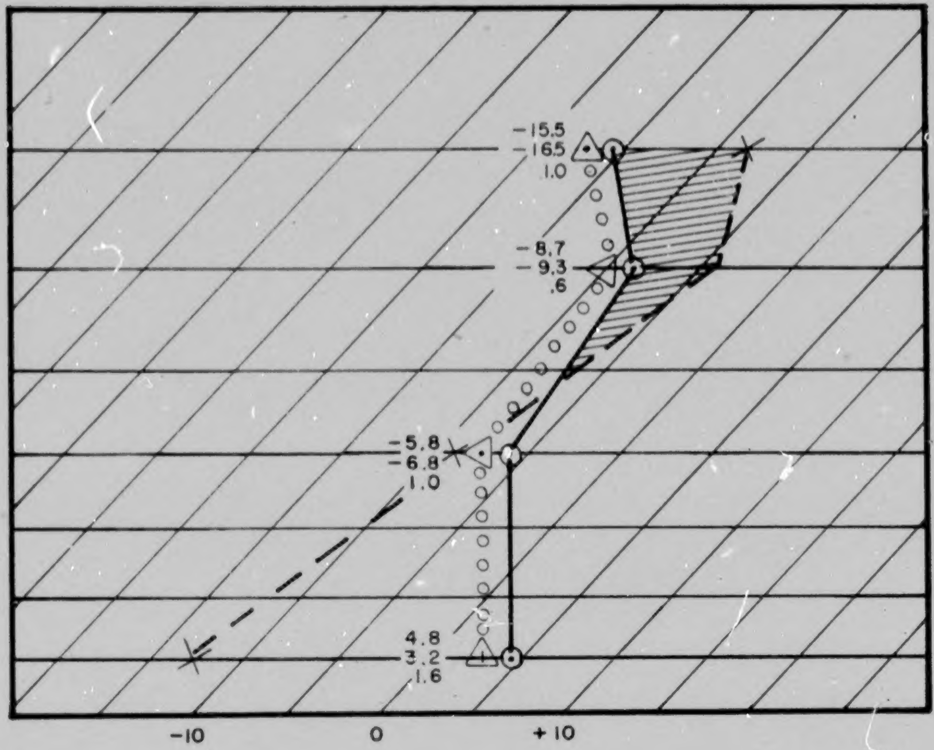

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Figure 11-26.—The −8D ice forecast method.

the lapse rate is shown by the same type comparison to be stable, and the clouds there should be altostratus.

Now determine the freezing level. It is noted that the lapse rate crosses the 0°C temperature line at 710 mb approximately 9,500 feet. Since it has been determined that clouds exist at this level, the temperatures are between 0° and −8°C, and the clouds are cumuliform, forecast clear ice in the unstable cloud up to 12,000 feet. Above 12,000 feet, the clouds are stratiform, so rime icing should be forecast. The intensity of the icing would have to be determined by the considerations given in the previous section of this chapter.

Forecasting Icing Using the −8D Method

When surface charts, upper air charts, synoptic and airway reports, and pilot reports are not clear as to presence and possibility of icing, it may be determined from Skew T by the following method:

1. Plot the temperature against pressure as determined from a raob sounding.
2. Write the temperature and dewpoint in degrees and tenths to the left of each plotted point.
3. Determine the difference (in degrees and tenths) between the temperature and dewpoint for each level. This difference is D, the dewpoint deficit; it is always taken to be positive.
4. Multiply D by −8 and plot the product (which is in degrees Celsius) opposite the corresponding temperature point at the appropriate place.
5. Connect the points plotted by step 4 with a dashed line in the manner illustrated in figure 11-26.

6. The icing layer is outlined by the area enclosed by the temperature curve on the left and the -8D curve on the right. In this outlined area, supersaturation with respect to ice exists. This is the hatched area as shown in figure 11-26.

7. The intensity of icing is indicated by the size of the area enclosed by the temperature curve and the -8D curve. In addition, the factors given in the following section should be considered when formulating the icing forecast. The cloud type and the precipitation observed at the raob time or the forecast time may be used to determine whether icing is rime or glaze.

The factors which were mentioned in the preceding paragraph which go into making the ice forecast are:

1. When the temperature and the dewpoint coincide in the raob sounding, the -8D curve must fall along the 0°C isotherm. In a subfreezing layer, the air would be saturated with respect to water and supersaturated with respect to ice. Light rime icing would occur in the altostratus-ninbostratus in such a region, and moderate rime icing would occur in cumulonimbus virga in such a region. Severe clear ice would occur in the stratocumulus virga, cumulus virga, and stratus.

2. When the temperature and dewpoint do not coincide but the temperature curve lies to the left of the -8D curve in the subfreezing layer, the layer is supersaturated with respect to ice and probably subsaturated with respect to cloud droplets. If the clouds in this layer are altostratus, altocumulus, cumulogenitus, or altocumulus virga, only light rime will be encountered. If the clouds are cirrus, cirrocumulus, cirrostratus, or cirrus nothus, only light hoarfrost will be sublimated on the aircraft. In cloudless regions, there will be no supercooled droplets, but hoarfrost will form on the aircraft through direct sublimation of water vapor.

3. When the temperature curve lies to the right of the -8D curve in a subfreezing layer, the layer is subsaturated with respect to both ice and water surface. No icing will occur in this region.

Forecasting Icing Using Graphs

Additional graphs and overlays for upper air diagrams have been prepared and are included in the Air Weather Service Publication, Forecaster's Guide on Aircraft Icing, AWSM-105-39. The diagrams may be used to construct locally prepared charts whereby icing conditions may be forecast from cloud types, temperatures, dewpoint spreads, and other information. The methods are much too elaborate and detailed to cover in this training manual. The reader is referred to the above publication for further information on this subject.

Modification of the Icing Forecast

The final phase is to modify the icing forecast as obtained by the foregoing methods. This is essentially a subjective process. The forecaster should consider the following items: probable intensification or weakening of synoptic features, such as low-pressure centers, fronts, and squall lines during the time interval between the latest data and the forecast time; local influences, such as geographic location, terrain features, and proximity to ocean coastlines or lake shores, radar weather observations; pilot reports of icing; and the like. The forecaster should be cautious in either underforecasting or overforecasting the amount and intensity of icing. An overforecast results in a reduced payload for the aircraft due to increased fuel load, while an underforecast may result in an operational emergency.

TURBULENCE

Turbulence is of major importance to pilots of all types of aircraft and therefore also to the Aerographer's Mate whose duty it is to recognize situations where turbulence may exist and to forecast for flight operations both the areas and intensity of the turbulence. In this chapter, the causative factors, classifications, intensities, and methods of forecasting turbulence will be discussed.

TURBULENCE CHARACTERISTICS

Turbulence may be defined as irregular and instantaneous motions of air which is made up of a number of small eddies that travel in the general air current. Atmospheric turbulence is caused by random fluctuations in the windflow. Given a smoothly analyzed wind field with both streamlines and isotachs smoothly drawn, any difference between an actual wind and this smooth field is attributed to turbulence.

To an aircraft in flight, the atmosphere is considered turbulent when irregular whirls or eddies of air affect the motion of the aircraft and a series of abrupt jolts or bumps is felt by the pilot. Although a large range of sizes of eddies exists in the atmosphere, those causing bumpiness are roughly of the same size as the aircraft dimensions and usually occur in an irregular sequence imparting sharp translation or angular motions to the aircraft. The intensity of the disturbances of the aircraft varies not only with the intensity of the irregular motions of the atmosphere but also with aircraft characteristics such as flight speed, weight stability, and size.

TYPES OF TURBULENCE

The atmosphere is always and everywhere turbulent, but often the intensity of turbulence is so small that it has a negligible effect on aircraft operations. Large intensities of turbulence are to be expected whenever the shear (horizontal or vertical) is large or wherever we find instability. Turbulence, for the purpose of this discussion, is divided into the following causative factors:

1. Thermal or convective. This type of turbulence is caused by localized vertical currents due to surface heating or unstable lapse rates and by cold air moving over warmer ground. It is more pronounced over land than water and in the daytime during summer.
2. Mechanical. Whenever wind speeds near the ground are high enough, shear becomes significant, causing small eddies and gusts. The degree of turbulence is proportional to the roughness of the surface and to the wind velocity. The higher the wind velocity and the rougher the terrain, the greater is the turbulence.
3. Frontal. This type turbulence results from the local lifting of warm air by cold air masses, or the abrupt wind shear (shift) associated with most cold fronts.
4. Large scale wind shear. Marked gradients in wind speed and/or direction, due to general variations in the temperature and pressure fields aloft, are the primary cause of this type turbulence.

Two or more of the causative factors often work together. In addition, turbulence is produced by "manmade" phenomena, such as in the wake of an aircraft.

PROPERTIES OF TURBULENCE

Turbulence Close to the Ground

Turbulence close to the ground is a combination of mechanical and convective turbulence. When the lapse rate is neutral or stable, conditions are simpler. In that case, only mechanical turbulence is of importance. This depends on the wind speed and the roughness of the ground.

In the case of mechanical turbulence close to the ground, the intensity decreases with increasing height. This is not the case when the mechanical turbulence is mixed with convection, as for example behind a cold front. In that case the convective clouds indicate rough flying up to considerable height.

Convection differs in two important features from mechanical turbulence: it is characterized by larger horizontal wavelengths, and it produces much stronger lateral fluctuations.

The intensity of convective turbulence generally increases with height, reaching a maximum in the upper half of convective clouds. This increase will be stopped by inversions.

To summarize, the most intense turbulence and gustiness near the ground occur under conditions of unstable lapse rates, strong winds, and rough ground.

Turbulence in Cumulus Clouds

The appearance of cumulus clouds makes the presence of a great deal of turbulence apparent. In general, the taller the cloud, the more turbulence there is. Although the initiating mechanism in cumulus type turbulence is thermal instability, much of the roughness is caused by smaller eddies produced by the large shears between up and down drafts.

Intermingling of downdrafts with the original updrafts results in severe turbulence, particularly at the boundary between regions of precipitation and regions of no precipitation. The same result is also confirmed by radar studies of thunderstorms made in this country and in England: severe turbulence occurs in regions of strong echoes, and particularly in regions of sharp boundaries of echoes. Hence, as has been proved recently, radar is an excellent aid for the avoidance of extremely turbulent conditions in flights through thunderstorms.

The intensity of turbulence in thunderstorms seems to increase with height well past the middle of the clouds. The Thunderstorm Project found the maximum effect of turbulence on aircraft occurred between 15,000 feet and 20,000 feet. The actual turbulent velocities may increase to even greater elevations, but have less effect on aircraft because the air density is lower at higher levels.

In the last stage of the thunderstorm, the air subsides and the storm dissipates without any indication of severe turbulence.

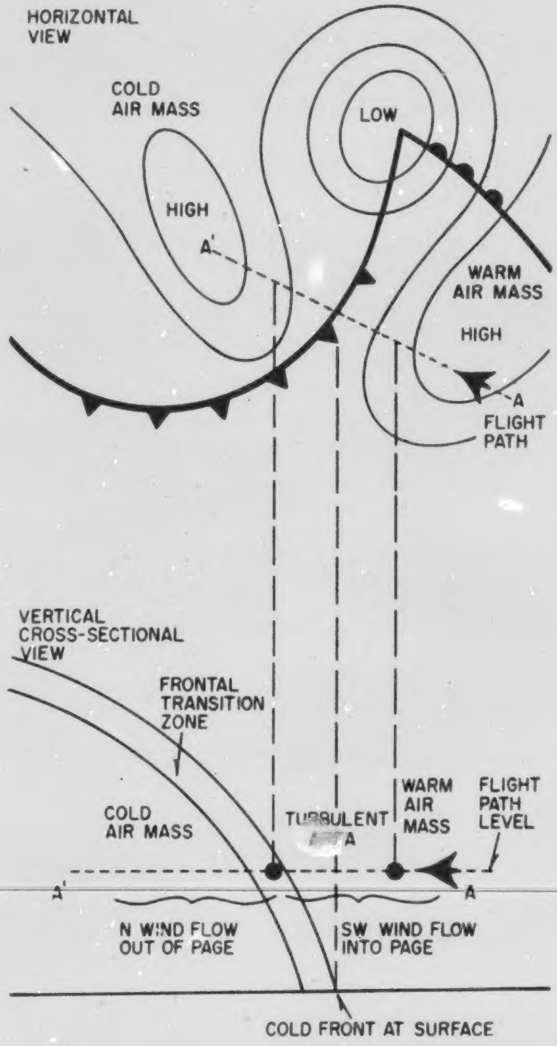
The intensity of turbulence in cumulus clouds depends on the difference between the temperatures inside and outside the cloud.

Present methods of estimating turbulence expected in cumulus type clouds before the clouds have formed generally make use of the relation between temperature differences.

Turbulence in the Vicinity of Fronts

Frontal turbulence is caused by the lifting of warm air by a frontal surface leading to instability and/or mixing or shear between the warm and cold air masses. The vertical currents in the warm air are strongest when the warm air is moist and unstable. The most severe cases of frontal turbulence are generally associated with fast moving cold fronts or squall lines. In these cases, mixing between two air masses as well as the differences in wind speed and/or direction (wind shear) add to the intensity of the turbulence.

Excluding the turbulence that would be encountered along the front, figure 11-27 illustrates the wind shift that contributes to the formation of turbulence across a typical cold front. As a general rule, the wind speed is stronger in the colder air mass.



AG.634

Figure 11-27.—Turbulence across a typical cold front.

Wind Shears and Clear Air Turbulence

A relatively steep gradient in wind velocity along a given line or direction (either vertical or horizontal) produces churning motions (eddies) which result in turbulence. The greater the change of wind speed and/or direction in the given direction, the more severe will be the turbulence. Turbulent flight conditions are often found in the vicinity of the jetstream where large shears in the horizontal and vertical are found. Since this type of turbulence may occur is perfectly clear air without any visual warning in the form of clouds, it is often referred to as clear air turbulence.

The term "clear air turbulence" is misleading because not all high level turbulence included in this classification occurs in clear air. However, the majority 75 percent is found in cloud-free atmosphere. Clear air turbulence is not necessarily limited to the vicinity of the jetstream, and may occur in isolated regions of the atmosphere. Most frequently, the clear air turbulence is associated with the jetstream and the mountain wave. However, it may also be associated with a closed low aloft, a sharp trough aloft, and an advancing cirrus shield. Too, a narrow zone of wind shear, with its accompanying turbulence, is sometimes encountered by aircrews as they climb or descend through a temperature inversion. Moderate turbulence may also be encountered momentarily when passing through the wake of another aircraft.

The criteria for each type of clear air turbulence (CAT) are:

1. Mountain wave CAT. Winds 25 knots or greater, normal to terrain barriers and the presence of significant surface pressure differences across such barriers.
2. Trough CAT. That portion of a trough which has horizontal vector shear across it of the magnitude of 25 knots in 90 nautical miles or greater.
3. Closed low aloft CAT. The circulation around the closed low aloft may be accompanied by the shears necessary to produce CAT. If the flow is merging or splitting, moderate or severe CAT can be encountered. Also to the northeast of a cutoff low aloft, significant CAT can be experienced. Just as with the jetstream CAT, the intensity of the turbulence is related to the strength of the shears.
4. Wind shear CAT. Those zones in space in which wind speeds are 60 knots or greater, and both horizontal and vertical shear exist according to the following criteria with intensities indicated in table 11-2.

No provision is made for light CAT because light turbulence serves only as a flight nuisance.

Table 11-2.—Wind sheer CAT with wind speed 60 knots or greater.

Horizontal shear (naut/mi)	Vertical shear (per 1,000 ft)	CAT intensity
25k/90	9-12k	Moderate.
25k/90	12-15k	Moderate, at times severe.
25k/90	above 15k	Severe.

Any of the above situations can produce moderate to severe clear air turbulence. However, the combination of two or more of the conditions is almost certain to produce severe or even extreme CAT. A jetstream may be combined with a mountain wave, or be associated with a merging or splitting low.

Turbulence on the Lee Side of Mountains

When strong winds blow approximately perpendicular to a mountain range, the resulting turbulence may be quite severe. Associated areas of steady updraft and downdraft may extend to heights from 2 to 20 times the height of the mountain peaks. Under these conditions when the air is stable, large waves tend to form on the lee side of the mountains and extend up to the lower stratosphere for a distance up to 100 miles or more downwind. They are referred to as standing waves or mountain waves and may or may not be accompanied by turbulence. Some pilots have reported that flow in these waves is often remarkably smooth while others have reported severe turbulence. The structure and characteristics of the mountain wave were presented in chapter 5 of this training manual. Refer back to the diagram in that chapter for a typical vertical illustration.

The windflow normal to the mountain produces a primary wave and generally less intense additional waves farther downwind. The characteristic cloud patterns may or may not be present to identify the wave. The pilot for the most part is concerned with the primary wave because of its more intense action and proximity to the high mountainous terrain. Severe turbulence frequently can be found out to 150 miles downwind, when the winds are greater than 50 knots at the mountaintop level. Moderate turbulence often can be experienced out to 300 miles under the previously stated conditions. When winds are less than 50 knots at the mountain peak level, a lesser degree of turbulence may be experienced.

Some of the most dangerous features of the mountain wave are the turbulence in and below the roll cloud, the downdrafts just to the lee side of the mountain peaks and to the lee side of the roll clouds. The cap cloud must be always avoided in flight because of turbulence and concealed mountain peaks.

These six rules listed have been suggested for flights over mountain ranges where waves exist:

1. The pilot should, if possible, fly around the area when wave conditions exist. If this is not feasible, he should fly at a level which is at least 50 percent higher than the height of the mountain range.
2. The pilot should avoid the roll clouds since these are the areas with the most intense turbulence.
3. The pilot should avoid the strong downdrafts on the lee side of the mountain.
4. He should also avoid high lenticular clouds, particularly if their edges are ragged.
5. The pressure altimeter may be as much as 1,000 feet off, near the mountain peaks.
6. The airspeeds recommended for the aircraft should be observed in penetration of the turbulent areas.

CLASSIFICATION AND INTENSITY OF TURBULENCE

COMNAVWEASERVCOM INSTRUCTION 3140.4 sets forth a common set of criteria describing the meteorological characteristics with which the respective classes of turbulence are typically associated in that the weather forecaster may evaluate and forecast degrees of atmospheric turbulence uniformly throughout the Naval Weather Service.

This instruction further states that the Gust Criteria shown in the Turbulence Criteria Table (table 11-3) be adopted as standard and that the idealized descriptions entitled Guide to Turbulence Classes as Typically Associated with Meteorological Conditions be adopted as a guide for meteorologists forecasting turbulence. Therefore, Naval Weather Service Units must adopt as standard the Gust Criteria, and utilize the guide to turbulence classes as a guide in forecasting turbulence.

Guide to Turbulence Classes

As typically associated with meteorological conditions, the following is a guide to classification of turbulence.

EXTREME TURBULENCE.–This rarely encountered condition is usually confined to the strongest forms of convection and wind shear, such as:

1. In mountain waves in or near the rotor cloud (or rotor action) usually found at low level leeward of the mountain ridge when the wind component normal to the ridge exceeds 50 knots near the ridge level.
2. In severe thunderstorms where available energy indicates the production of large hail (three-fourths inch or more), strong radar echo gradients or almost continuous lightning. It is more frequently encountered in organized squall lines than in isolated thunderstorms.

SEVERE TURBULENCE.–In addition to the situations where extreme turbulence is found, severe turbulence may also be found:

1. In mountain waves:

   a. When the wind component normal to the ridge exceeds 50 knots near the ridge level: at the tropopause up to 150 miles leeward of the ridge. (A reasonable mountain wave turbulence layer is about 5,000 feet thick.)

   b. When the wind component normal to the ridge is 25-50 knots near the ridge level: up to 50 miles leeward of the ridge, from the ridge level up to several thousand feet above and at the base of relatively stable layers below the tropopause.

Table 11-3.—Turbulence criteria

		TRANSPORT AIRCRAFT OPERATIONAL CRITERIA[2]		GUST CRITERIA
Adjectival class	Airframe limits[1]	Descriptive	Air Speed fluctuation	Derived gust velocities (U_{de})[3] the order of
Light	Not specified	A turbulent condition during which occupants may be required to use seat belts, but objects in the aircraft remain at rest.	5 to 15 knots	5 to 20 fps
Moderate	Not specified	A turbulent condition in which occupants require seat belts and occasionally are thrown against the belt. Unsecured objects in the aircraft move about.	15 to 25 knots	20 to 35 fps
Severe	Not specified	A turbulent condition in which the aircraft momentarily may be out of control. Occupants are thrown violently against the belt and back into the seat. Objects not secured in the aircraft are tossed about.	More than 25 knots.	35 to 50 fps
Extreme	a. Positive and negative gusts greater than 50 fps (U_{de}) at V_C[4] between sea level and 20,000 ft for transport category aircraft. b. Positive and negative gusts greater than 30 fps (U_e)[3] at all speeds up to V_C for normal, acrobatic, and utility aircraft.	A rarely encountered turbulent condition in which the aircraft is violently tossed about, and is practically impossible to control. May cause structural damage.	rapid fluctuations in excess of 25 knots	more than 50 fps

Table 11-3.—Turbulence criteria—Continued.

Footnotes: [1] As derived from the Flight Loads section CAM 4b, Airplane Airworthiness, Transport Categories (May 1960); and CAM 3 Airplane Airworthiness: Normal, Utility, and Acrobatic Categories (Nov. 1959) of Civil Air Regulations.

[2] Aircraft Turbulence Criteria developed by NACA Subcommittee on Meteorological Problems (May 1957).

[3] U_e approximately equals 3/5 U_{de}.

[4] V_C is the design cruising speed.

2. In and near mature thunderstorms and occasionally in towering cumuliform clouds.
3. Near jetstreams within layers characterized by horizontal wind shears greater than 16 knots/degree latitude (40 knots/150 nautical miles) and vertical wind shears in excess of 6 knots/1,000 feet. When such layers exist favored locations are below and/or above the jet core and from roughly the vertical axis of the jet core to about 50 or 100 miles toward the cold side.

MODERATE TURBULENCE.—In addition to the situations where extreme and severe turbulence are found, moderate turbulence may also be found:

1. In mountain waves:
   a. When the wind component normal to the ridge exceeds 50 knots near the ridge level: between the surface and about 10,000 feet above the tropopause from the ridge line to as much as 300 miles leeward.
   b. When the wind component normal to the ridge is 25-50 knots near the ridge level: between the surface and the tropopause from the ridge line to as much as 150 miles leeward.
2. In, near, and above thunderstorms and in towering cumuliform clouds.
3. Near jetstreams and in upper trough, cold, low, and front aloft situations where vertical wind shears exceed 6 knots/1,000 feet or horizontal wind shears exceed 7 knots per 1 degree latitude.
4. At low altitude (usually below 5,000 feet above the surface) when surface winds exceed 25 knots or the atmosphere is unstable because of strong insolation or cold advection.

LIGHT TURBULENCE.—In addition to the situations where more intense classes of turbulence occur, the relatively common class of light turbulence may be found:

1. In mountainous areas even with light winds.
2. In and near cumulus clouds.
3. Near the tropophase.
4. At low altitudes when winds are under 15 knots or where the air is colder than the underlying surface.

FORECASTING TURBULENCE CLOSE TO THE GROUND

Over land at nighttime there is very little turbulence close to the ground. The only exception is the case of high wind speeds over rough terrain. This kind of turbulence decreases with increasing height.

During the day turbulence close to the ground depends on the radiation intensity, the lapse rate, and the wind speed. Turbulent intensities tend to increase with height throughout the unstable and neutral layers above the ground up to the first inversion or stable layer. Similar turbulence occurs in fresh polar outbreaks over warm waters.

Vertical gustiness increases with height more rapidly than horizontal gustiness.

Situations for particularly violent turbulence near the ground occur shortly after cold front passages, expecially over rough ground. Other examples of rough air are conditions over deserts on hot days and thunderstorms.

Peak gusts at the surface can be estimated to be essentially equal to the wind at gradient level (except in thunderstorms).

FORECASTING TURBULENCE IN CONVECTIVE CLOUDS

In this section, methods for determining the type and intensity of turbulence in convective clouds after the characteristics of the upper air sounding have been determined are discussed.

Eastern Airlines Method

In the absence of any dynamic influences which might serve to drastically modify the vertical temperature and moisture distribution, radiosonde data on hand may be used to evaluate the turbulence potential of convective clouds or thunderstorms for periods up to 24 hours. The following is one procedure for predicting turbulence in such clouds. The material used in this section is reproduced by permission of the Academic Press, from the book, Weather Forecasting for Aeronautics, by J. J. George and Associates. Such permission is gratefully acknowledged.

The technique is as follows:

1. Determine the CCL.
2. From the CCL, proceed along the moist adiabat to the 400-mb level. This is the updraft curve.
3. Compare the departure of the updraft curve with the free air temperature curve and note that value of maximum positive departure up to 400-mb. For positive values, the updraft curve should be warmer than the free air temperature curve. The value of the maximum positive departure obtained in this step is referred to as ΔT.

Table 11-4 is based on several years relating ΔT values to commercial pilot reports of thunderstorm turbulence, and can be used to predict the degree of turbulence in air mass thunderstorms.

Table 11-4.—Relation of maximum positive departure to thunderstorm turbulence.

ΔT (°C)	Turbulence
0-3	Light
4-6	Moderate
7-9	Severe
Above 9	Extreme

This method of forecasting thunderstorm turbulence is almost exclusively confined to the warmer months when frontal cyclonic activity is at a minimum. During the cooler months, frontal and cyclonic influences may cause rapid changes in the vertical distribution of temperatures and moisture and some other methods have to be used.

Other Methods

Various other methods utilized in forecasting turbulence have been developed. A National Weather Service method employing an overlay on the plotted upper air sounding diagram is one. The Air Force uses a modification of the Eastern Airlines Method in which the atmosphere is divided into two layers—surface to 9,000 feet MSL and above 9,000 feet MSL.

These methods are discussed, with examples, in NAVAER 50-1P-546, An Introduction to Atmospheric Turbulence. Forecasters should refer to this publication for further understanding of them.

FORECASTING CLEAR AIR TURBULENCE

Clear air turbulence (CAT) poses as one of the most comm in-flight hazards encountered by the modern high altitude, high performance aircraft.

Not all high level turbulence occurs in clear air. However, the rough cobblestone type of bumpiness does occur in clear air or the cloudless portion of the sky, without visual warning. The turbulence may be violent enough to disrupt tactical operations and possibly cause serious aircraft stress.

Most cases of clear air turbulence at high altitudes can be associated with the jetstream or more specifically with abrupt vertical wind shear. They are experienced most frequently during the winter months when the jetstream winds are the strongest.

The association of clear air turbulence with recognizable synoptic features meets only with limited success. However, using the following as general areas where CAT may occur, flight crews should be briefed as to the possibility of its occurrence:

1. In general, in any region along the jetstream axis where wind shear appears to be strong horizontally, vertically, or both.
2. In the vicinity of the traveling jet maxima, particularly on the cyclonic side; most occurrences of moderate and severe clear air turbulence have been reported in this area.
3. In the jetstream front, below, and to the south of the core.
4. Near 35,000 feet in cold deep troughs.

Mountains wave turbulence should be forecast if:

1. The flow at and above mountaintop level is approximately constant in direction and nearly at right angles to a mountain barrier having a steep lee slope.
2. The wind at mountaintop level exceeds 25 knots.
3. There is a rapid increase in wind speed with altitude in the level of the mountain range and for several thousand feet above of at least 30 knots for weak wave development, and at least 50 knots for strong wave development. (A peak in the vertical wind profile near or somewhat above mountaintop level is a characteristic of strong waves.)
4. An upwind inversion or stable layer exists near mountaintop height.

Observation of Vertical Wind Shear by Radar

When precipitation is showery, the cells may be seen on the RHI scope as separate columns. These columns are often distorted from the true vertical by wind shear. To observe this shear effect clearly, the antenna must be scanning in the same plane as the wind shear. To do this, adjust the azimuth until the greatest distortion from the vertical is observed. Such observations give heights of shear zones and a qualitative estimate of the shear.

Stability, Instability, and Turbulence by Radar

Radar can be useful in a qualitative evaluation of the degree of turbulence near the terminal in those layers where precipitation is forming or through which it is falling. Horizontally stratified echoes are indicative of smooth air; whereas, vertical columns or cellular echoes indicate vertical motions which cause turbulence to aircraft. In addition, the sharpness of the bright band is an indication of the instability involved. Sharp wind-shear layers are also an indication of associated turbulence.

It is well known that the most severe turbulence (as well as heavy icing and damaging hail) is associated with actively developing thunderstorms extending to great heights. During this dangerous growing stage, the top of the radar echo rises rapidly. The growth can best be followed by using the RHI scope and scanning vertically on the azimuth which includes the highest echo. Subsidence of the top of the echo indicates the end of convection and the resulting decrease of turbulence in that particular cell or group of cells.

CONDENSATION TRAILS

A condensation trail (contrail) is a visible trail of small water droplets or ice crystals formed under certain conditions in the wake of an aircraft.

The formation of contrails is considered to decrease significantly the effectiveness of operational aircraft under combat conditions. In daylight when clear weather prevails, the enemy's detection problem is practically solved when the aircraft produces contrails. At night, contrails greatly simplify the enemy's interception problem.

To remove the detection hazard during combat missions, it is necessary to be able to predict the altitudes at which contrail formation is probable so that flight at those altitudes may be avoided. Such prediction is the function of the weather office which provides the briefing for the flight.

TYPES OF CONTRAILS

There are three types of contrails: aerodynamic, instability, and engine exhaust. The aerodynamic contrail is produced by the momentary reduction of pressure resulting from the flow of air past an airfoil. It is of short duration, and for this reason is not considered an operational hazard. The instability contrail is produced by the passage of an aircraft through an otherwise undisturbed layer of unstable air with a higher relative humidity. Conditions conducive to such formations exist only rarely. The most prevalent of contrails is the engine exhaust contrail, and it is with this type that you as an Aerographer's Mate will be concerned, since it is the only one which must be forecast.

FORMATION OF CONTRAILS

When an aircraft passes through the atmosphere, the engine releases a certain quantity of water vapor and heat as a result of combustion processes in the engine. The exhaust mixes with the air. The water vapor introduced tends to increase the relative humidity and to bring the air closer to saturation. The heat released tends to decrease the relative humidity. The formation of a contrail depends, therefore, upon such factors as the amount of heat and water produced by the fuel combustion, the wake or entrainment characteristics of the aircraft, and the original temperature and humidity condition of the undisturbed air.

Entrainment

Entrained air is that which is drawn into and mixed with the exhaust gases of the aircraft. The amount of air entrained into the exhaust trail varies continuously from near zero immediately behind the aircraft to an extremely large amount far behind it. The rate at which air is entrained varies with the type of aircraft, power setting, speed, and the density and stability of the atmosphere. The ratio of entrained air to exhaust gas, at a given distance behind tha aircraft, is greatest when the aircraft is operating at its most efficient speed and altitude.

Meteorological Factors

If the temperature, pressure, and humidity are suitable, the water vapor of the exhaust may produce supersaturation with consequent contrail formation. The more humid the air at a given temperature and pressure, the greater will be the tendency for contrails to form.

NOTE: Only in the case of jet aircraft can a definite relationship be established between pressure, temperature, and relative humidity. In propeller-driven aircraft, energy losses are variable. Since all the energy is not contributed to the wake, only an estimate of the limiting temperatures for contrail formation can be given.

CONTRAIL FORECASTING

Calculations based on the rates of heat dissipation and mixing with entrained air permit the determination of a maximum temperature for each pressure and relative humidity value above which contrails cannot form in the wake of a jet aircraft.

The Forecasting of Aircraft Condensation Trails, AWSM 105-100, provides all the basic background for producing contrail forecasts for both propeller driven and jet aircraft.

The Skew T, Log P diagrams (DOD-WPC 9-16 and 9-16-2) are overprinted with a set of lines that represent theoretical critical relative-humidity values separating the categories for forecasters to utilize in determining the probability of, or absence of, contrails from jet aircraft. The application of these curves is fully explained in Air Weather Service Manual 105-100.

CHAPTER 12

TROPICAL ANALYSIS AND FORECASTING

The importance of tropical analysis and forecasting has been reemphasized in recent years as U. S. operations in low latitudes have increased. Various aspects of tropical analysis and forecasting that were poorly founded or lacking in entirety have been brought to light. Concerted effort has been made to improve the analysis and forecasting techniques to provide more informative and accurate forecasts to users.

Also during recent years there has been a great increase in the amount of observational data available to the forecaster. This has been provided by meteorological satellites and weather radars, as well as an increased number of weather reporting stations.

During any discussion of the Tropics, the actual area under consideration is questionable; therefore, for the purpose of this chapter, the Tropics are defined as the region lying between the high pressure belts at the surface of each hemisphere, and the troposphere and lower stratosphere above this region. Therefore, the boundaries of the Tropics will vary with space and time. They may approach within 15° latitude of the equator or recede to as much as 45° latitude from it.

Since many Navy weather units are located or operating in the Tropics, it is of the utmost importance that the principles of analysis and forecasting be understood.

In the temperate and frigid zones, the Aerographer's Mate is required to acquire a knowledge of the air masses and the boundaries between them in order to effect a valid analysis and the subsequent forecast. To forecast in the Tropics, on the other hand, the Aerographer's Mate finds that he is dealing with what may properly be termed a single air mass, the air in the equatorial zone. Since the contrasts between masses of air in convergent or divergent flow in relation to each other are minute, his primary concern is with internal changes in this vast belt of equatorial air.

Success in tropical analysis and forecasting, much like analysis in the Temperate Zones, will depend largely on the experience of the Aerographer's Mate. The experience of the Aerographer's Mate as an analyst in this case implies and includes a thorough knowledge of the climatology of the analysis area.

BASIC PRINCIPLES OF TROPICAL FORECASTING

With the advent of satellite systems, tropical forecasting has improved very much. The combined use of extensive climatology and satellite information will produce the best forecasts possible in the tropical areas.

WEATHER VARIATIONS WITHIN THE TROPICS

Most of the area within the tropics is oceanic. However, since this region contains mountainous islands, coastal areas, and large portions of some continents, some discussion must be given to the variations in tropical weather in each of these situations.

Island and Coastal Tropical Weather

During the day as warm, moist air moves inland and is lifted over the terrain, large cumuliform clouds develop. These clouds are common in coastal areas. The lifting of moist air on the windward side of mountainous islands also produces towering cumulus clouds which may frequently be seen from long distances, indicating the presence of an island ahead. Low islands also cause more clouds and rain than over the water, but the clouds tend to increase downwind to a maximum near the lee end of the island. Cumuliform clouds and precipitation are more abundant over island and coastal areas than over the open oceans, except in the vicinity of certain tropical weather circulation systems, such as hurricanes, typhoons, the intertropical convergence zone, and tropical waves.

Continental Tropical Weather

The greatest temperature and pressure variations in the Tropics are found in continental areas. The temperature on continents undergoes a significant daily temperature variation, becoming very warm in the afternoon and becoming cool during the night.

Cloudiness and precipitation are at a maximum for the Tropics over land areas which are exposed to airflow from ocean areas. This helps account for tropical jungle areas.

Oceanic Tropical Weather

Typical weather over the tropical oceans is characterized by cumuliform clouds, although all forms of clouds are observed. Although most oceanic cumulus clouds have a common base of approximately 2,000 feet, occasionally lowering to 1,500 feet or 1,000 feet in precipitation, there is great variation in the heights of the tops, which depends primarily on the area in which they form. Doldrum cumulus, which form in the areas of light winds and small wind shear, have tops varying from 6,000 to 12,000 feet.

The term "trade cumulus" applies to cumulus clouds that develop in the broad easterly windstream between latitudes 10 and 30 degrees. Their chief characteristic is the inclination of the cloud axis caused by wind shear through the cloud layer. The height of the trade cumulus tops varies from about 7,000 to 9,000 feet. Tall cumulus clouds occur only at widely scattered intervals in the trade wind. Seen from the air, the low trade cumulus have a characteristic arrangement in bands paralleling the windflow. Scattered rain showers are common from these cumulus clouds and visibilities are good except in the showers. Over tropical oceans, cumulonimbus are almost always restricted to areas affected by synoptic or meso-scale disturbances; tops range from 30,000 to 40,000 feet, except in areas of intense tropical storms where they may extend beyond 60,000 feet.

The tropical oceanic regions are characterized by a mean temperature near 80°F throughout the year. The mean air temperature is generally within a few degrees of the sea-surface temperature (SST) except near some coasts where there is a prevailing offshore flow of cold continental air.

Wind is the most important factor in tropical weather, even over open ocean areas. When there is convergence of windflow, there is a piling up of air locally. This lifting action in warm, moist air results in cumuliform clouds and precipitation. These cumuliform clouds often develop to great heights, and heavy showers are frequently observed.

WEATHER ELEMENTS IN THE TROPICS

Winds

Two levels that are important for analyzing and forecasting synoptic-scale tropical weather systems are one near the surface and one in the upper troposphere. Fortunately the gradient level and the 200-mb level are suited to provide this data and they are also levels where there is an abundance of data. The gradient level is defined as the lowest level at which predominately friction free flow occurs. Over most of the tropics this is taken to be 3,000 feet MSL; however, in some cases this may have to be adjusted upward due to land elevation.

The major features shown by the mean sea-level pressure fields (subtropical high-pressure ridges and low-latitude troughs) dictate the variations of the wind flow within the tropical region.

Two basic types of circulation patterns are evident in the low-latitude trough zone. These types are called the monsoon and trade wind troughs. The monsoon troughs occur near the large continental areas due to the land-sea monsoonal effects and are characterized by a directional shear zone with westerlies on the equatorward side and easterlies on the poleward side. The trade wind troughs occur primarily over the oceanic areas of the North Atlantic, Northeast and North Central Pacific and are characterized by confluence of trade wind flows from the northern and southern hemispheres.

Other types of circulation patterns also occur in the near equatorial region in various locations. One of these circulation features is a wind system near the equator where the trade wind flow from one hemisphere changes direction as it moves into the opposite hemisphere. This is called a buffer zone; it separates two wind streams of opposing directions, the easterly trades and the monsoon westerlies. The sense of the rotation in the buffer zone varies from clockwise during the northern hemisphere summer to counterclockwise during the winter.

Other prominent circulation features on the gradient-level charts are the subtropical ridges with their associated anticyclonic cells and the heat lows over the subtropical continental areas.

Detailed information on wind flow both at the low and upper levels is contained in AWS Technical Report 240, Forecasters Guide to Tropical Meteorology.

Temperature Distribution

In general, the horizontal temperature distribution over the Tropics can be summarized as follows: The annual temperature range in the Tropics does not vary on the average more than 1° to 3°C. The range is in the neighborhood of 5° to 10°C over land and only 2° to 3°C over the ocean. The diurnal range is much greater than the annual range and is much greater over land than water masses. The largest variations take place in the fringe areas of the Tropics which come under the influence of mid-latitude pressure systems. The eastern parts of oceans are colder than the western parts due to the circulation around the subtropical highs which brings midlatitude colder air into these areas. Temperature, therefore, is one of the least important elements on the tropical synoptic surface map.

One very significant feature of the vertical temperature distribution of the Tropics is the presence of very dry warm air aloft. This air is the result of subsidence in the subtropical anticyclones. It is therefore dry aloft and has a temperature inversion in the lower levels. This inversion of temperature is called the trade inversion. It is generally located below 10,000 feet. This temperature inversion is more pronounced over the eastern parts of the oceans due in part to cooling below from cold ocean currents. Over the west coasts of continents fog and stratus are common. Below this inversion the lapse rate is very steep, and above the inversion the lapse rate closely approximates that of the dry adiabatic lapse rate. This low inversion over the eastern portions of oceans undergoes a transition westward and completely disappears over the western portions of the oceans, though stable layers may appear in this region in specific weather patterns.

Inversions over the eastern portions of the oceans are formed by a broad-scale descent of air from higher altitudes in the eastern end of the subtropical highs. The eastern side of the high may be thought of as tilting downward and the western end upward. Therefore, the mass of air moving eastward on this side tends to subside and to be lifted as it moves westward on the southern side of the cell. As this air aloft descends, it meets opposition from the low-level maritime air flowing equatorward. The height which the inversion forms depends on the depth to which the upper air current is able to penetrate downward. The inversion is considered to be a mixing zone between these two masses.

The height of the trade inversion has a significant effect on tropical weather. When the base of the inversion is at a high altitude, the moist layer underneath has a greater thickness and clouds build to greater heights, coverage is more extensive and rain is more likely. When the base of the trade inversion is low in altitude, moisture is confined to lower levels, cloud cover and tops are lower, and rain is less likely. Weather is better when the trade inversion is low except for periods of reduced visibility due to haze and over the cold water belts near the west coasts of continents, where fog and stratus are common.

A common phenomenon of the trade winds is the presence of trade wind cumulus. This occurs in the absence of large-scale convection or orographic lift produced by islands in the daytime, which may give rise to towering cumulus and abundant showery type rainfall.

From one-third to one-half of the tropical ocean area is covered by this type of clouds.

The seasonal range of temperatures aloft below 300 millibars is usually in the 1° to 3°C range. Cooling generally takes place below 700 millibars. Over the western portions of the Pacific ocean the monsoon season both winter and summer greatly affects the temperatures, moisture content, and stability of the air masses.

Moisture Distribution

Humidity is almost constant throughout the moist layer except at the base and lower part of the cloud layer where a slight maximum is found. The average depth of the moist layer is between 7,000 to 8,000 feet above which it decreases sharply. This decrease is coincident with the temperature inversion. Although the temperature inversion is a good indication of this moisture break, the break is evident in some cases where this inversion is not present.

Cloud Types

A popular conception of the cloud distribution in the Tropics pictures the equatorial regions as a tremendous factory continuously producing cumulonimbus type clouds which because of the height of the tropopause rise to spectacular heights. THIS IS TRUE ONLY OF CERTAIN REGIONS. All types of high, middle, and low clouds are present in tropical regions. Although records indicate that cumulonimbus type clouds are more common in tropical than in polar regions, they also show that except for such places as central Africa, southeast Asia, Indonesia, the Amazon Valley, and the southern United States, cumulonimbus is the exceptional cloud, rather than the most common. In fact, certain equatorial regions are known to report few, if any, cumulonimbus type clouds throughout the year. Cumulus, however, in one form or another is the predominant form of tropical cloud.

The climatology of tropical clouds like those in middle and high latitudes may be divided into two parts: Climatology of clouds over oceans and climatology of clouds over land. The orographic effect of islands upon the type and form of cloud cover is even more pronounced in the topical oceanic areas than elsewhere. The higher mean temperatures of tropical air, with its greater capacity to hold water vapor, results in orographic clouds with a minimum amount of forced lifting. Another factor governing oceanic cloudiness in the Tropics is the strength and height of the trade inversion. As was pointed out earlier, the moisture content of the air above the trade inversion decreases markedly, and strong stability at the inversion effectively limits the vertical extent and development of clouds. Too, in the Tropics, the absence of a variety of air masses of different origin eliminates the association between fronts and clouds so pronounced in higher latitudes and leaves us only with the basic differences between air-mass clouds over land and those over water surfaces.

CIRRIFORM CLOUDS.—Cirriform clouds in varying amounts are found everywhere in the oceanic Tropics. Isolated maximums of high cloudiness are found near the Equator, some of which may be attributed to the prevalance of the anvil tops of cumulonimbus. An overall seasonal variation in high cloudiness is not evident in the Tropics. Certain areas do experience seasonal changes, but they are not large enough to be considered representative of the Tropics as a whole.

MIDDLE CLOUDS.—Middle clouds are likely to be found everywhere in the Tropics and in any season of the year; no appreciable seasonal variation occurs when taking global distribution into consideration; and no pronounced minimum or maximum of middle cloudiness is evident along any particular latitude in the Tropics.

LOW CLOUDS.—Cumulus is the predominant type of cloud found in the Tropics. Their development, vertical and horizontal extent, and persistence are controlled by several factors: horizontal convergence in the wind field, depth of the moist layer, orography, and vertical stability of the air mass. There are many significant types of cumulus and they usually range in some intermediate form between cumulus humilis and cumulus congestus. Stratus and stratocumulus are also found in certain regions of the Tropics.

Satellite cloud pictures have been studied to determine the tropical large-scale organization of cloud systems in various regions of the tropics. Three major types of cloud systems were identified and given names. These are cloud clusters, monsoon clusters, and popcorn cumulonimbi. All cloud systems are composed of cumulonimbus type clouds. More detailed information concerning the significant features of each cloud system is found in AWS Technical Report 240.

Of all forms of cloudiness in the Tropics, the low cloud still commands the greater portion of the forecaster's attention. In many tropical areas, an increase or decrease or change in the form of low cloudiness is often an indication that the NORMAL weather and cloud pattern will be disturbed.

Precipitation

The general belief that frequent and heavy precipitation is one of the major characteristics of the Tropics is only partly true. While it is a fact that the Tropics as a whole receive more rainfall than other portions of the earth, the climatological records also show that the annual amount of rainfall within its boundaries varies tremendously both in time and space. These variations can be as much as 300 inches annually between two stations 70 miles apart.

The zone of maximum rainfall in the Tropics lies along the equatorial trough. The minimums lie along the usual position of the subtropical anticyclonic belt.

The Tropics may be considered to consist of three basic rainfall regimes:

1. In the vicinity of the equatorial trough, characterized by rain at all seasons of the year.
2. In the trades, characterized by summer rains and dry winters.
3. In the subtropical anticyclonic belt, characterized by dryness at all seasons of the year.

These regimes give only a very broad indication of the zonal distribution of total rainfall. Specifically, within each region, the rainfall in any year may or may not fall within the scheme of this classification.

The predominant form of precipitation in the Tropics is of the showery type. However, when tropical disturbances are present, rain may be steady or intermittent from the associated altostratus deck or low stratiform clouds.

GENERAL ASPECTS OF TROPICAL ANALYSIS

There are major differences in techniques utilized for tropical analysis as compared to those used in the temperate zones. This provides the tropical analyst and forecaster with a new challenge.

The air mass concept as applied in the mid-latitudes does not generally hold true in the tropics, especially on a day to day basis. Although some writers have made a distinction in the tropical air masses such as a monsoon air mass, an equatorial air mass, and a tropical air mass, in actual practice it is almost impossible to find any boundaries between them to which the criteria of a front can be applied. The whole tropical troposphere with a few local exceptions, fulfills the standard definition of a single air mass. The Aerographer's Mate is chiefly concerned with internal changes within this air mass.

In order to observe these changes a more complete method of recording elements at selected stations is necessary than the normal 6-hour synoptic map. This evaluation should be done by graphically recording various elements hourly. The elements covered should include pressure, temperature, dewpoint, visibility, weather, cloud amount and type, and wind.

Elements which ordinarily are conservative in high latitudes, that is, change very slowly with time during the transformation of an air mass, may change more rapidly in the Tropics. Moreover, changes of these elements in the temperate zone may be attributable to only one cause, while a very different cause may operate in the Tropics. Thus changes in dewpoint are usually slow within a single temperate-zone air mass and are attributable to the influence of the surface over which the air mass is moving. In low latitudes, certain rapid changes in dewpoint can occur within the tropical air mass which have little or nothing to do with the surface over which the air mass is moving.

For these reasons, the Aerographer's Mate who is undertaking analysis in the Tropics has to take special care in evaluating the data, at least until he has become thoroughly familiar with the geography of the region and with the often surprising local effects characteristic of this zone.

The analyst in the tropics must be aware that local effects often outweigh the synoptic changes that are occurring, especially near the surface. The local effects are caused primarily by radiational cooling and heating, and topographical features, as well as convective activity at or near the station during observational periods. If these effects are overlooked or ignored, they may lead to errors in data interpretation as well as analysis. Therefore it is incumbent on the tropical analyst to evaluate the synoptic data as well as become thoroughly familiar with topographical features of stations to be able to separate local effects from the actual synoptic changes that are occurring.

Sizable departures from the normal should be closely checked when they appear in reports from tropical stations. This may be attributed to observational errors or due to some feature that has gone unnoticed until this time. A number of tropical meteorologists consider climatology as the best method of forecasting within the tropics due to the fact that most changes in the weather are not rapid.

DATA CONSIDERATIONS

To insure that all available data are plotted and are as accurate as possible, an analyst, first and foremost, must use his managerial skill and initiative to see that observations are not lost through ignorance or neglect by the map plotters. At a weather office, the plotting personnel must be trained constantly to be on the alert; the analysts must be able to spot check all work.

An analyst must learn which stations can be relied upon to transmit accurate weather information. A little experience at a tropical analysis center will show quickly that some stations send much better reports than others. The reports judged reliable should be plotted and adhered to rigorously. This applies to both surface and upper air data.

In tropical weather forecasting the forecaster must be well acquainted with the local situation of all stations that furnish the basis for the analysis. As an aid, weather offices should keep a running file on all key stations, the topographic conditions, other local peculiarities, the diurnal course of weather, and the pressure and wind in each month. For each of the 6-hourly standard observation periods, this information could be posted in map report form near the analysis desk for ready reference. Additional statistical data, such as frequency distribution of pressure at a given time of day, could also be kept on file for reference. Also, obtain local area Forecaster's Handbooks for as many stations as possible.

The main considerations of the weather elements with reference to tropical analysis and forecasting are presented in the following sections.

Surface Data

SURFACE PRESSURE. Sea level pressures and pressure changes, in conjunction with the gradient wind pattern, furnish a basic tool of analysis in low as in high latitudes. Pressures are far more numerous than upper winds or raobs; therefore, the maximum information must be extracted from the pressure data.

In any given month the range of observed pressures will be almost wholly within a 10-mb spread in a given tropical area. Deviations from normal greater than 0.5 millibar are rare. It follows that the deviation of a pressure from the normal must be known to the nearest 0.5 millibar in order to be of much value in tracing departure patterns with any accuracy. This then is the upper limit permissible for observers' and other errors.

Except in the vicinity of a tropical storm, 3-hourly tendencies are nearly useless in the Tropics and can never be exploited as in middle latitudes except as they deviate from normal. There are three reasons for this:

1. The 3-hour synoptic pressure change is so small (order of 0.1 millibar) as to fall within the unavoidable range of observation errors.
2. The diurnal variation (order of 1 millibar) completely masks the true synoptic variations.

3. Passage of local cloud systems often affects the barometer to 0.1 millibar or more.

Consequently, the 24-hour pressure change is used in tropical analysis. This eliminates the normal diurnal change from the tendency and also provides for a better relation between the rate of motion of disturbances and the time interval over which pressure changes are measured. The 24-hour change in the Tropics is comparable to the 12-hour change in middle latitudes.

In most areas, changes of 3 millibars are rare and definitely indicate danger of a severe development when a storm is not already in existence. Even values of 1.5 to 2.5 millibars warrant careful attention, especially when the change is at or below average pressure values.

Pressure gradients in tropical regions have the order of 1 millibar in 100 miles, and less. This means that most of the character of a map is lost if 4-mb isobars are drawn. Isobars should be drawn for every 2 millibars south of the subtropical ridge and for 1 millibar within 5° of the meteorological Equator. It also follows that the pressure must be known within at least 0.5 millibars; or else severe unrealistic distortion of isobars will regularly occur.

There are certain topographic effects which cause true pressure abnormalities. In particular there are dynamic pressure reductions on the lee side of mountain ranges and in channels between mountainous islands. These reductions may amount to 1 to 3 millibars. Unless this is recognized and taken into account, the permanent cyclonic deformations of isobars that appear in certain areas on most charts will not be interpreted correctly.

SURFACE WIND.–In middle latitudes, surface winds of 7 to 16 knots or stronger are quite representative, except in mountainous country, at coasts, or in thunderstorms. In the Tropics, topographic and coastal features assume even greater importance in producing diurnal wind regimes. They may render the surface wind completely unrepresentative, but if care is used, they can usually be deduced by subtracting local effects. The situation as encountered over large flat land areas and over ocean areas is quite different. Ship winds are reliable; they can usually be accepted at face value. As in middle latitude, except winds observed in and near heavy showers, the daytime wind over flat stretches of land is most representative.

Since the greatest percentage of tropical land station winds comes from areas subject to marked diurnal changes, the question arises whether anything at all can be done with these winds. As in the case of pressure, departures from normal and 24-hour changes provide one possible route.

In hurricane forecasting, one looks for weak or westerly surface winds where normally easterlies of 7 to 16 knots should be blowing. This technique is sometimes applicable. One could, in fact, draw different vectors between the normal and observed wind. This has not been attempted on a quantitative basis, but it has been used as a mental aid. Qualitatively one may note, for instance, whether the strength of the easterlies is above or below normal, or if the winds at a station are becoming more southerly with time. Here again the time series representation is a major aid; the observed time changes of wind must be correlated with those of pressure and other elements.

TEMPERATURE AND DEWPOINT.–In tropical regions, temperature and dewpoint are not of any great value to an Aerographer's Mate for forecast purposes. The reason for this is that the temperature will fluctuate quite rapidly with the passage of local showers over both land and ocean areas. Rain evaporating during its fall causes both temperature and dewpoint to approach the wet-bulb temperature. Thus readings in showers are quite unrepresentative. Even when a shower has ended, the low-level air cannot immediately return to its previous state.

Exceptions occur near the boundaries of continents. Here the air-mass differences between continental and oceanic air may be as great as anywhere in middle latitudes. Dewpoints lower than average also will be found in regions with marked suppression of the normal convection over oceans, and on the lee side of islands.

At this time temperature and dewpoint can be used mainly to estimate low cloud bases. Since these cannot be computed closer than 200 to 400 feet, fluctuations of 1°F at the surface do not detract from the validity of the calculation.

CURRENT AND PAST WEATHER.—Over most of the Tropics, the reports of current weather nearly always involve some form of precipitation. However, in some areas, for instance the Sahara and Northern Australian deserts, dust storms are more important, especially in the dry season. Along the cold water coasts of Africa and America, fog takes first place in importance.

Although it is highly desirable to separate rainfall into showery and steady precipitation, it is not always done. In addition, it is not always easy to make this observation. A station may receive fairly steady rain for a long time; yet the rain is derived from cumuliform clouds. Because of the cell character of tropical rain, and again because of local regimes, the current weather report is very often rather meaningless. Many occasions are on record when severely disturbed conditions happened to let up temporarily at the 6-hourly observation period. Here a check of hourly sequences, when available, is most helpful. Again, in some freak situation, stations have been deluged when all around them conditions were quite normal. Although not all of these phenomena can be spotted, it is advisable to place as much (or more) weight on weather in the past 6 hours as on current weather.

It is hoped that with time a place for radar rain data will be found in the synoptic codes. A single radar set scans a very wide area. Such a view is vastly superior to all spot reports of weather.

STATE OF SKY.—A good picture of the distribution of cloud types and amounts, in conjunction with present and past weather, is in a sense the core of the analysis problem. Unfortunately, cloud reports often are one of the least satisfactory items in the surface report. Usually clouds cannot be measured but must be estimated. The present codes do not permit an observer to describe properly the various states of sky found in low latitudes. The observer's entry nearly always entails some arbitrary decision. Some observers will regularly transmit the same combination of low, middle, and high clouds. One may find that at neighboring stations entirely different types are reported on a routine basis, especially middle and high clouds.

In spite of these drawbacks much can be extracted from the cloud data. Emphasis must be placed on the presence or absence of low, middle, or high clouds as groups. Division into one of nine types within each group should generally be disregarded, especially with reference to middle and high clouds. Although there are many regional and seasonal variations, the following three cloud combinations have been found to occur most frequently over the west Atlantic area in the hurricane season:

1. With disturbed conditions, frequent cumulus congestus, and cumulonimbus with layers of middle and high clouds.
2. With suppressed convection, small amounts of cumulus, sometimes larger amounts of very shallow cumulus, almost no stratocumulus, and no middle or high clouds.
3. Intermediate cumulus of average height, isolated cumulus congestus, and cirrus in varying amounts.

The first two types indicate definite synoptic conditions, and all other information should be considered with this in mind. Although the middle and high cloud types unfortunately remain vague, the presence or absence of such clouds is highly significant.

Concerning representativeness for synoptic purposes, ship data can be taken at face value to the extent that the observations are considered reliable.

VISIBILITY.—Visibility in the Tropics is usually good. It is a minor element in the analysis compared to middle latitudes.

Here we need to consider only haze, which over the oceans is mostly produced by accumulation of salt particles absorbing moisture. This is called wet haze. Such accumulation goes with very stable conditions and suppressed convection, often in conjunction with strong surface winds. Many salt particles are picked up from the sea; these, however, cannot penetrate the strong inversion lid present, nor are they precipitated back to the ocean due to the lack of showers. Since a definite synoptic condition is indicated, it is worthwhile to keep track of intensity and extent of wet haze.

Dry haze occurs occasionally when continental air with a high dust content moves over the ocean. This permits assigning a source region for the air. Although of not much interest in general forecasting, this may be of importance for special purposes. Haze also is produced at times following volcanic eruptions when a large part of the erupted material is trapped under a trade inversion. Such haze has been known to persist for more than a week, gradually spreading over a large area.

Upper Air Data

UPPER WINDS. Pilot balloon and rawin data generally are representative except when wind speeds are less than 5 knots. The main thing to ascertain is the quality of upper wind soundings. At times, upper winds plotted on time sections at certain stations yield a fantastic sequence. Wind directions and speeds change violently with height and time, though the weather remains the same and no other station shows such remarkable fluctuations. When this happens, it is best to separate the good from the bad. Never discard a whole report, because part of it is usually good. You just have to find it.

Here again, with experience the Aerographer's Mate will learn which stations can be relied upon to transmit accurate wind data.

UPPER AIR SOUNDINGS. Certain information is provided with considerable accuracy by radiosonde data. Temperature inversions, especially the large trade wind inversions, are recorded very satisfactorily. The same holds true for the low-level moisture distribution.

Comparisons between careful measurements made by specially equipped aircraft and raob data have brought out that a dry adiabatic lapse rate is normal for the subcloud layer but that the radiosonde often shows a more stable structure. In the high troposphere, observers frequently encode only the required standard levels and do not transmit enough significant points. As a result the lapse rate shows changes at each standard level, which is obviously not true. Tropopause pressures are also affected by this practice, even though many observers are apt to enter one significant point for very well-pronounced tropopauses rather than put them at one of the standard levels.

Upper air soundings can be most unreliable in the Tropics at certain times; therefore care should be used before a sounding is accepted as valid. See if it fits the synoptic situation. Temperature and moisture distribution will vary considerably if one instrument ascends just outside a thunderstorm or shower, and the precipitation is of a local nature and is unrepresentative of the surrounding area.

Reconnaissance Reports

Regularly scheduled reconnaissance reports by weather squadron aircraft and unscheduled reports by other aircraft supplement the surface and upper air reports received from land stations and ships. Flight procedures change from year to year and are fully described in other publications.

CENTER FIX. It is very dangerous procedure to disregard a reported position of the storm center which does not agree with your preconceived ideas. There have been a few erroneous reports by reconnaissance; therefore, proceed cautiously when faced with the above situation. Over the years, these standards of accuracy have been set: dead reckoning fix, within 30 miles; Loran fix in a single line area, within 15 miles; and Loran fix in a good area, within 10 miles. It is worthwhile to follow successive positions closely, correlate them with wind reports, and see that the latest reported aircraft position fits with the capabilities of the aircraft in use. This serves, also, as a check on the storm center position when obtained. Radar fixes will normally fall within the limits of accuracy set forth above. However, it is a good plan to obtain a series of fixes before making a final decision. If the storm eye is not clearly defined, radar fixes have a disconcerting habit of jumping around. They need some averaging.

When frequent fixes are given, sometimes at intervals as small as 20 minutes, the inherent errors in navigation and determination of the exact center of an eye may lead the unwary forecaster into embarrassing traps. Do some reasonable smoothing and keep in mind that there are minor oscillations of the storm center about the mean track.

WINDS. Surface wind reports from reconnaissance aircraft able to observe the sea are quite reliable. The error in the reported direction will usually be under 10°. For speed the error will not exceed 5 knots when the wind is below 60 knots, and it will be under 10 to 20

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knots at speeds from 60 to 100 knots. Flight-level winds will be about as accurate in an area where Loran facilities are available.

Most authorities disagree as to how accurate the speeds are above 100 knots, but it should be assumed the error will be much larger. In considering the accuracy of the surface wind, consideration must be given to the height of the aircraft. The accuracy will be less reliable if the aircraft is above 10,000 feet, and if there are many clouds present, the reported wind may not be the highest, but will be the highest observed.

Flight-level winds obtained with the Doppler equipment are very accurate and compare closely with rawin reports.

Other flight-level winds are averaged over a given distance and should be considered carefully especially where Loran facilities are poor. The best way to judge these types of reports is to have several reports in a given area and use an average of all the reports.

The wind report should be carefully plotted with a protractor on a large-scale map. Direction and speed should be entered by number along with wind barb. The relative accuracy of the wind reports merits careful treatment, as (1) they are used for center-position determinations in the event the flight aborts and does not penetrate far enough for a center fix, (2) they will enter into computations of objective forecasting techniques, and (3) they are of value for research purposes.

OTHER ELEMENTS. Surface pressure reports are usually quite accurate, and heights of standard upper levels are good but may be in error at times. A reported height can be compared to a reporting rawinsonde station; if there is a difference, the difference is usually applicable to all the reports from the same aircraft.

The other elements such as weather, state of the sea, turbulence, clouds, temperature, and the like are all important and should be judged in the same manner as other similar data.

Commercial and Military Transport Aircraft Reports

These reports, when available, are of value. The present weather given in the reports is helpful, and the D-values are very useful in determining the wind field. Accuracy of D-values will vary radically with the method of position determination and due to the limitations of instruments. The D-value and gradient between consecutive reports from the same aircraft are usually quite accurate. Wind reports are averaged over a period of 20 to 30 minutes. This is a point often overlooked when analyzing the data.

If you have access to the navigators of these flights, it is possible to conduct a successful program for improvement in the caliber of reporting. This is an opportunity which should not be missed.

ANALYSIS APPROACH

A rational approach to analysis in any region depends on (1) the objectives of the analysis, (2) the type and quality of the observations, and (3) the space and time distribution of the observations. So far, we have discussed point (2). With the preceding evaluation of the reports in mind, consider now point (3) and several steps which will permit an analyst to come as closely as possible to a representation of the space and time distribution of pressure, temperature, wind, and weather.

Both the lower and the higher troposphere must be considered in forecasting. Thus, you need good upper air charts as a basic step toward good prognoses. The data, however, are scattered, often so widely that whole synoptic systems can be situated between stations and escape notice. Make as much of the upper air data at these scattered stations as possible; this is accomplished with the time-section technique. Further, utilize the more numerous and frequent surface observations to help with the upper air analysis. Do this with the technique known as differential analysis.

In preparing charts, you must work both downward from the upper air information and upward from the surface reports.

The analysis procedure is broken down into four successive steps in the remaining sections of this chapter. They are as follows:

1. Complete utilization of the upper air data with the time section method.
2. Low-level wind analysis. Pilot balloons in the lower troposphere are far more frequent

than high-reaching soundings. Since the wind field at the gradient level is usually a better indicator of low-level conditions than surface isobars, this step defines the low-level disturbances and helps guide the drawing of isobars.

3. Surface analysis. It consists of three parts: pressure, pressure change, and cloud and weather analysis.

4. Preparation of contour and contour change charts at upper isobaric surfaces with the differential analysis method.

APPLICATION OF SATELLITE DATA

If one area where satellite data has been the greatest asset to the meteorologist were to be singled out, more than likely it would be the Tropical region.

The great ocean expanses, as well as sparsity reports, within this region have enabled weather phenomena to develop to great proportions without being discovered in normal analysis.

With the advent of satellite pictures on a continuing basis, the meteorologist has been provided with a tool which, when used properly, will aid in discovery, as well as improve the understanding of the various tropical phenomena.

Interpretation of the satellite pictures for the tropical phenomena will be discussed in that portion of this chapter dealing with tropical analysis.

TROPICAL PHENOMENA

Within this section of the chapter we will discuss phenomena that are found within the tropical region and in most cases have no corresponding mid-latitude phenomena. Among those phenomena discussed here will be the shear lines, subtropical jets, tropical easterly jet, tropical waves, vortices, the intertropical convergence zone (ITCZ), and tropical cyclones. For more in-depth discussion of these and other tropical phenomena it is recommended that you refer to AWS Technical Report 240, Forecasters Guide to Tropical Meteorology.

SHEAR LINE

A shear line is defined as a line or narrow zone across which there is an abrupt change in the horizontal wind component parallel to this line. It most commonly refers to lines of cyclonic shear. Monsoon and upper tropospheric troughs as well as remnants of old cold fronts are examples of shear lines.

It has been known for many years that cold fronts from mid-latitude penetrate deep into the tropics and occasionally move across the equator. With the meteorological satellite pictures available today it has become relatively easy to track the cold fronts (shear lines) well into the tropics, over oceanic areas. The leading edge of the front is usually marked by a pronounced line of convection and a series of convection lines oriented parallel to the front (and to the wind) may occur on the poleward side of the main line. The average tops in these shear lines are usually not high (10,000 to 15,000 feet) but the associated low ceilings and rainfall along the line may cause poor terminal weather conditions, especially with orographic effects.

Surface cold fronts often penetrate to low latitudes over tropical continental areas during the cold season. Even though currents of polar air are usually rather shallow by the time they reach subtropical latitudes, a surface temperature and dewpoint discontinuity, and a shear line, can be maintained to low latitudes in continental areas because of the repeated nocturnal radiational cooling in the clear, dry air mass behind the front.

SUBTROPICAL JET STREAM (STJ)

Subtropical jet streams are a persistent feature of the tropical general circulation. A detailed study of the northern hemisphere subtropical jet stream during winter shows it to be a simple broadscale current with very high wind speeds. The subtropical jet stream is continuous around the world and speeds of 150 to 200 knots are not uncommon. There is a basic three wave pattern with ridges, and maximum wind speeds are over the east coasts of Asia and North America and the Middle East. The mean latitude is 27.5N ranging from 20 to 35 degrees North. Averaged around the hemisphere, the core speed is about 140 knots, located near the 200-mb level.

TROPICAL EASTERLY JET (TEJ)

The tropical easterly jet is a persistent feature over extreme southern Asia and northern Africa during the northern hemisphere summer. The tropical easterly jet extends over the layer from 200- to 100-mb in the latitude belt 5 to 20 degrees North. It is remarkably persistent in its position, direction and intensity.

TROPICAL WAVES

A tropical wave is defined as a trough or cyclonic curvature maximum in the trade wind easterlies. The wave may reach maximum amplitude in the low or middle troposphere or may be the reflection from the upper troposphere of a cold low or equatorward extension of a mid-latitude trough.

First extensive investigation into tropical waves was begun in the 1940's. Using surface pressure data they were able to track the movement of isallobaric centers across the western Atlantic and Caribbean Sea area. They found these isallobaric centers were accompanied by westward moving, wave-like oscillations in the basic easterly current in the lower troposphere. These waves moved at average speeds of 10 to 15 knots, reached their maximum intensity in the layer from 700 to 500 mb, and sloped eastward with height. In the typical case where the wave moved slower than the basic current in lower levels, and faster than the basic current in upper levels, the area west of the wave trough was characterized by subsidence and fair weather, while areas of convergence and disturbed weather occurred east of the trough. The basic tropical wave model is shown in figure 12-1. The top portion of the figure shows an east-west vertical cross section of the winds through the wave at its latitude of maximum development, and the bottom shows the surface pressure and 10,000 to 15,000 foot streamline patterns. The positions of the surface and upper level troughs and the weather distribution accompanying the wave are also indicated.

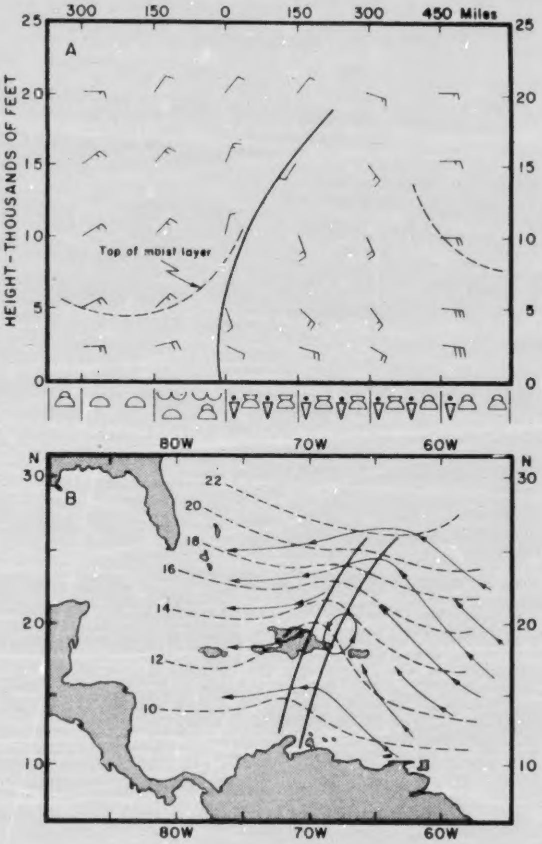

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Figure 12-1.—Model of a tropical wave both vertical and horizontal.

Later investigations revealed that a number of waves move westward faster than the basic current. In this case, convergence occurred west of the wave axis and became strong near the axis. The cloud pattern across the wave varied considerably, with cumuliform clouds including cumulonimbus evident west and a nimbostratus layer dominant east of the axis.

Another variation of cloud pattern associated with tropical waves was labeled the inverted-V formation. This name was adopted because the cloud bands are arranged in a herringbone pattern somewhat resembling a nested series of upside-down V's. Figure 12-2 illustrates this type of tropical wave. This type of wave is best defined in the eastern and central North Atlantic where it shows reasonably good day to day continuity. These waves move westward at an

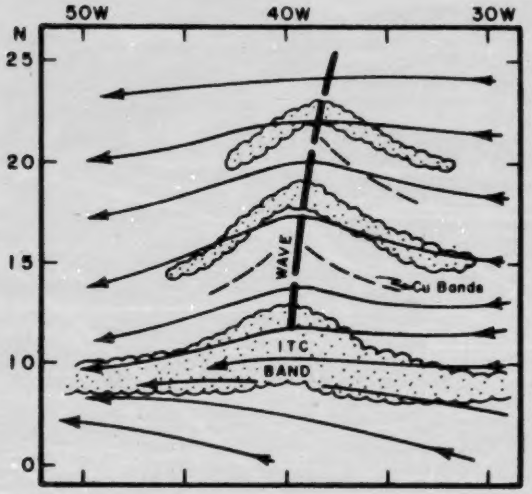

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Figure 12-2.—Inverted V tropical wave cloud pattern.

average speed of 16 knots which is about the speed of the low-level trades.

With the increased use of satellite data as well as increased amounts of meteorological data it has become apparent that there are a wide variety of weather producing systems in the tropics and that the classical tropical wave does not occur as frequently as originally believed.

VORTICES

Cyclonic cloud and circulation patterns occur frequently in the tropical region. These features are referred to as vortices.

Some of the weak vortices develop tropical storm or higher classification while others eventually dissipate with little intensification.

These vortices had, for the most part, gone undetected with the conventional reporting networks but became quite apparent as the use of satellite pictures became widespread.

It is still difficult, from the satellite photographs alone, to determine at what tropospheric level the circulation associated with the cloud vortex is most intense the level where the maximum cyclonic vorticity occurs.

To determine this, tropical meteorologists need to use all available conventional data and have a thorough knowledge of the climatology of various types of vortices in specific areas and seasons. A thorough and comprehensive discussion of this is given in AWS Technical Report 240, Forecasters Guide to Tropical Meteorology.

INTERTROPICAL CONVERGENCE ZONE (ITCZ)

As satellite pictures have greatly increased the amount of information provided meteorologists to improve forecasts, they have also led to some changes in theories concerning various phenomena. One of the phenomena so affected is Intertropical Convergence Zone (ITCZ), or as it is frequently referred to now, zone of Intertropical Confluence (ITC).

By definition the Regional Tropical Analysis Center denotes this as: "A nearly continuous fluence line (usually confluent) representing the principal asymptote of the Equatorial Trough." In general this refers to the area where horizontal convergence of the airflow is occurring.

Figure 12-3 shows a typical cloud band associated with the ITC. It is within this cloud band that disturbances frequently occur. The cloud band, at times, is narrow (2 to 3 degrees latitude) and continuous for thousands of miles. At other times, it is discontinuous and is characterized by a number of large cloud areas 5 to 10 degrees in latitude across. On occasion, vortical cloud patterns are observed within the ITCZ cloud band.

Generally, disturbances along the ITCZ move from east to west and can move poleward and develop into tropical storms. These disturbances are most frequent in the doldrum portions of the equatorial trough. In that area, low-level cyclonic wind shear is present over large areas. This, together with friction, produce the forced convergence necessary for development of the individual cloud systems which form the ITCZ cloud band. Figure 12-4 shows the cloud pattern typical of an active doldrum trough in the Western Pacific.

It has been determined that the presence of surges of flow from one hemisphere to the other is one of the controlling factors of ITCZ clouds. As air moves across the equator, anticyclogenesis

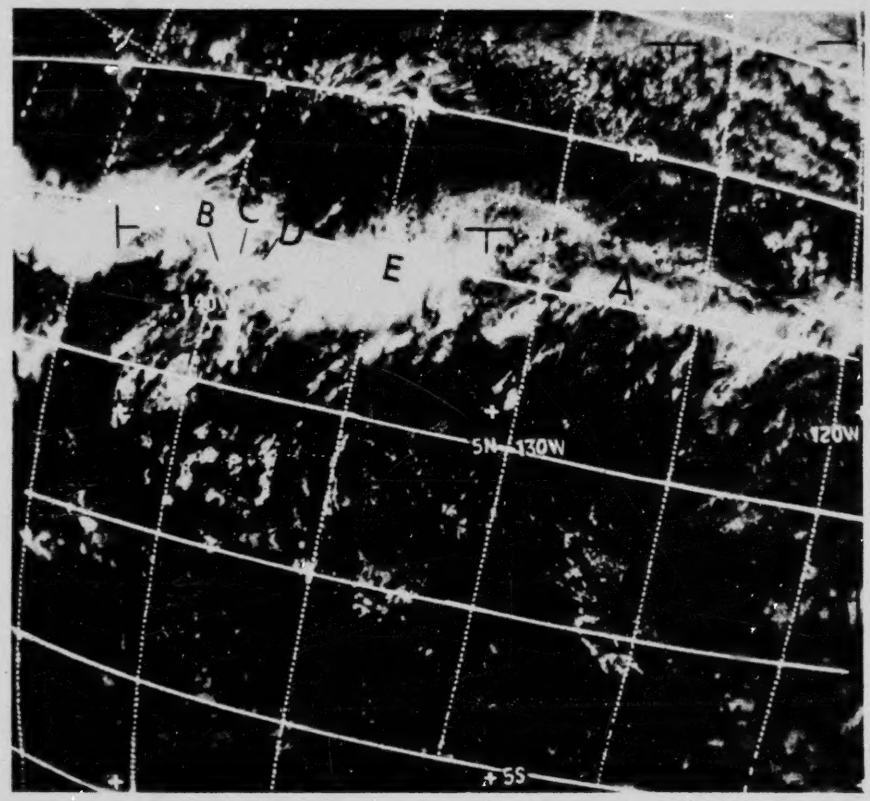

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Figure 12-3.—Satellite photograph of cloud band associated with ITCZ.

takes place. This results in a reduction or clearing of clouds along one portion of the ITCZ cloud band and an intensification of the cloud band in advance of the burst of cross equatorial flow.

Weather Along the ITCZ

The degree and severity of the weather along the ITCZ vary considerably with the degree of convergence between the two air currents. The zone of disturbed weather may be as little as 20 to 30 miles in width or as much as 300 miles. Under typical conditions, frequent rainstorms, cumulus and cumulonimbus type clouds and local thunderstorms occur. Violent turbulence may be associated within these storms. Cloud bases may lower to below 1,000 feet, or even be indistinguishable, in heavy showers. Their tops frequently exceed 40,000 feet. An extensive layer of altocumulus and altostratus usually occurs due to the spreading out of the upper parts of the clouds. These clouds vary in height and thickness with the currents of the air masses. At higher levels a broad deck of cirrostratus spreads out on both sides of the zone. Visibility is generally good except when reduced by heavy rain shower activity.

Surface wind in the vicinity of the ITCZ is generally squally in the heavy shower areas.

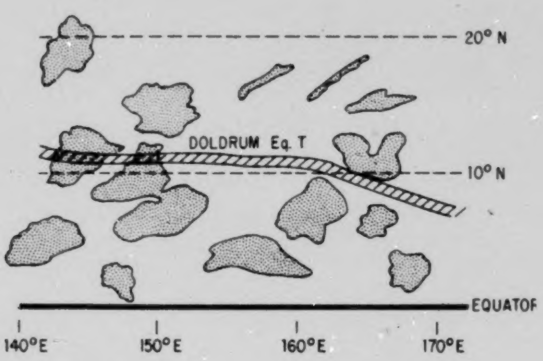
20° N
DOLDRUM Eq. T
10° N
EQUATOR
140°E
150°E
160°E
170°E

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Figure 12-4.–Typical satellite-observed cloud cluster patterns relative to a doldrum equatorial trough.

Usually these winds in the squalls do not exceed 15 to 25 knots but winds of 40 to 50 knots or higher have been reported.

Ice formation in the heavy cloud masses associated with the ITCZ is likely to reach serious proportions when pilots are flying at altitudes above 15,000 feet. (This is roughly the average freezing level in equatorial regions.)

The intensity of the ITCZ varies interdiurnally, from day to day and to a lesser degree annually. Over ocean areas, precipitation reaches its maximum just before dawn with a minimum occurring late in the morning or early afternoon. Over land areas the reverse is true, except on coastal areas when the wind has an onshore component. In this case the diurnal maximum of precipitation takes place in the early morning.

Seasonal Variation

Through the use of mapped digital satellite data for 1967, the seasonal meridional displacement of the cloud band associated with the ITCZ has been determined. See figure 12-5.

The cloud pictures revealed that the cloud band has somewhat different characteristics in different parts of the world. In the Atlantic, it is centered about 3N in the winter season and moves to about 8N by late summer. In the Pacific, the seasonal fluctuation of the cloud band is not readily apparent in that part of the band east of 150W. Seasonal pressure changes over North America may be responsible for the seasonal shift of the ITCZ cloud band in the eastern part of the Pacific Ocean. There is some evidence of a second cloud band associated with the ITCZ in the eastern Pacific. Besides the main band north of the equator, a weak band appears at 5S in the January through March average. The strength of the second band varies from year to year; in some instances it fails to develop at all.

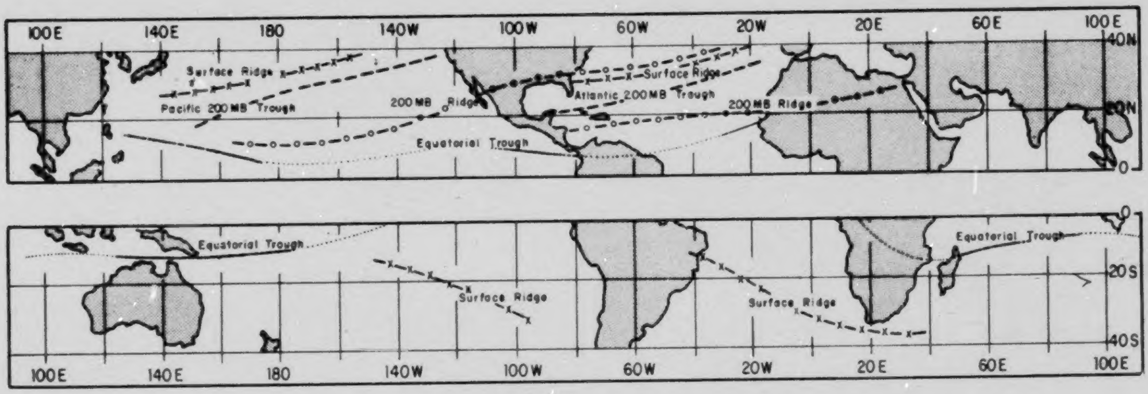
100E
140E
180
140W
100W
60W
20W
20E
60E
100E
40N
20N
0
Surface Ridge
Pacific 200MB Trough
200MB Ridge
Surface Ridge
Atlantic 200MB Trough
200MB Ridge
Equatorial Trough
0
Equatorial Trough
Surface Ridge
Surface Ridge
Equatorial Trough
20S
40S
100E
140E
180
140W
100W
60W
20W
20E
60E
100E

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Figure 12-5.–Mean summertime position of the ITCZ. (A) Northern Hemisphere; (B) Southern Hemisphere.

The ITCZ cloud band in the Indian Ocean during the Southern Hemisphere summer is much broader than those of either the Atlantic or eastern Pacific Oceans.

TROPICAL CYCLONES

The most destructive of all weather phenomena is the tropical cyclones. While a tornado exceeds the severity of a full-fledged tropical cyclone in a smaller area, it has a comparatively short path and life duration. The tropical cyclone due to its greater horizontal extent and longer life exceeds any other phenomena in total damage and loss of life.

Tropical cyclones have been given various names in different regions of the world, however they all have essentially the same characteristics. Although the tropical cyclone normally covers a large area, roughly circular or elliptical in shape, it is usually of small size compared with large extratropical cyclones. It differs also in that there are no distinct cold and warm sectors and no well-defined surfaces of discontinuity or fronts at the surface while the cyclone is in the Tropics. The tropical cyclone is found most frequently in summer or autumn of the hemisphere in which it occurs, while the extratropical cyclone is most frequent in cold months. Tropical cyclones have no moving anticyclones as companions while in the Tropics. In many other features, such as the region of calm or relative calm called the "eye," the east to west component of progressive motion in its early history, and the distribution of rainfall, tropical cyclones are distinctive. However, on leaving the Tropics they take on some of the characteristics of extratropical cyclones.

Classification of Tropical Cyclones

The nomenclature of tropical cyclones varies considerably. At times such terms as "tropical cyclone," "tropical storm," "hurricanes," and "typhoons" are used almost interchangeably with little regard for differences in size or intensity. There are generally three recognized categories of nonfrontal cyclones of tropical origin, all of which must show evidence of a closed circulation at the surface. These are distinguished in terms of observed or estimated surface wind speeds associated with the systems as follows:

1. Tropical Depression. Maximum winds less than 34 knots and one or more closed isobars on the surface. Normally these are expected to intensify.
2. Tropical Storm. Closed surface isobars with maximum winds 34 to 63 knots, inclusive.
3. Hurricane/Typhoon. Maximum winds of 64 knots or higher.

Life Cycle of the Hurricane/Typhoon

The energy that sustains tropical cyclones is derived from the energy that is released through the latent heat of condensation. The energy source is furnished to the tropical storm by the warm water sources over which it develops and moves. The warm moist air is lifted by a combination of convergence and instability of the air until it condenses. Upon condensation the latent heat is liberated. However, if the storm passes over a large land-mass, the source of energy is cut off and the storm will eventually dissipate. As the storm moves from southerly latitudes to higher latitudes, the moisture and heat source are no longer present and the storm will assume extratropical characteristics.

The average lifespan of this type storm is about 6 days from the time they form until they either move over a land surface or recurve to higher latitudes. Some storms last only a few hours, while some last as long as 2 weeks. The evolution of the average storm from birth to dissipation has been divided into four stages:

1. Formative or Incipient Stage. This stage starts with the birth of the circulation and ends at the time that hurricane/typhoon intensity is reached. This stage can be slow, requiring days for a weak cyclonic circulation to begin, or in the case of development on an easterly wave, it can be relatively explosive, producing a well-formed eye in as little as 12 hours. In this stage the minimum pressure reached is about 1,000 millibars. A good indication that a system of this type has formed or is forming is the appearance of westerly winds (usually 10 knots or more) in low tropical latitudes where easterly winds normally prevail.

2. Immature or Intensification Stage. This stage lasts from the time the system reaches hurricane-typhoon intensity until the time it reaches its maximum intensity in winds and its lowest central pressure. The lowest central pressure often drops well below 1,000 millibars and the wind system becomes organized in a tight ring around the eye with a fair degree of symmetry. The cloud and precipitation fields develop into narrow, inward spiraling bands. Usually the radius of the strongest winds are no more than 60 miles around the center. This development may take place gradually or occur in less than 1 day.

3. Mature stage. This stage lasts from the time the hurricane/typhoon attains its maximum intensity until it weakens to below this intensity or transforms to an extratropical cyclone. In this stage, the storm may exist for several days at nearly the same level of intensity or decrease slowly. The storm grows in size, with strong winds reaching farther and farther from the center. The weather and winds usually extend farther in the right semicircle of the storm. By the time the storm reaches this stage it is usually well advanced toward the north and west, or it has already recurved into the westerlies. The typhoons of the Pacific usually last longer in the mature stage and grow to larger sizes than hurricanes in the Atlantic.

4. Decaying Stage. This stage may be characterized by rapid decay as in the case of many storms which move inland, or after recurvature the transformation into a middle latitude cyclone. In the former case the storm steadily loses strength and character. In cases of transformation there is frequently a regeneration in the middle latitudes which results in maintenance or redevelopment of strong winds and other hurricane/typhoon characteristics.

There is no set duration for the time a storm may be in any one stage. It is entirely possible that a storm will skip one stage or go through it in such a short time that it is not distinguishable without the available synoptic data. Satellite picture interpretation has improved the possibility of observing the various stages of tropical cyclones. It has led to a newer classification of storms, from satellite pictures, which is divided into stages and types.

Characteristics of Hurricanes/Typhoons

To make the most efficient analysis of available data in the vicinity of tropical cyclones, the forecaster must be familiar with the normal wind, pressure, temperature, clouds, and weather patterns associated with these storms. No two tropical cyclones are exactly alike. On the contrary there are very great variations between storms. However, certain general features appear with sufficient frequency to predominate in the mean patterns. These features serve as a valuable guide when reconstructing the picture of an individual tropical cyclone from sparse data.

Since the meteorological elements are not distributed uniformly throughout all sections of the storm, it is customary to describe the storms in terms of four quadrants or right and left semicircles, separated by the line along which the center of the storm is moving and normal to this line at the cyclone center. Actually, in nature there are no clearly defined lines of demarcations between these areas. The general features of hurricanes/typhoons given in the following section apply mainly to the mature stage.

1. Surface Winds. The surface winds blow inward in a counterclockwise direction toward the center with those in the left rear quadrant having the greatest angle of inflow (in the Northern Hemisphere). The diameter of the area affected by the hurricane/typhoon force winds may be in excess of 100 miles in large storms or as small as 25 to 35 miles. Gale winds sometimes cover an area 500 to 800 miles or more. The maximum extent of strong winds is usually in the direction of the major subtropical anticyclone which is most frequently to the right of the line of movement of the storm in the Northern Hemisphere. No accurate measurement of the peak wind speeds of large mature storms has been made with any reliable degree of accuracy. Speeds of 140 knots have been recorded and it seems reasonable to assume that speeds of 200 knots may be attained at altitudes several hundred feet above the surface.

2. Surface Pressure. Since the central surface pressure of the mature storm is well below

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average, sea level isobars furnish an excellent tool for analysis of these storms. Isobars may be nearly symmetrical or they may assume an elliptical shape. The pressure gradient to the right of the storm's motion is strongest due to the forward motion of the storm against the existing pressure gradient. Various other deformations may occur, such as the extention of a trough southward from the storm. Barometric tendencies are not a particularly good indication of the movement of the storm outside of its sphere of influence. Usually the pressure falls with extreme rapidity in the 3 hours before the arrival of the storm and rises at an equal rate after passage. Central pressures of 950 to 960 millibars are not uncommon.

3. Surface Temperature. In contrast to extratropical cyclones the tropical cyclone may show no, or very little, surface temperature reductions toward the center of the storm, indicating that the horizontal adiabatic cooling due to the pressure reduction is largely offset by the addition of the heat through the release of heat to the atmosphere through the condensation processes. Upper air temperatures, it has been found, are actually higher by 5°C or more.

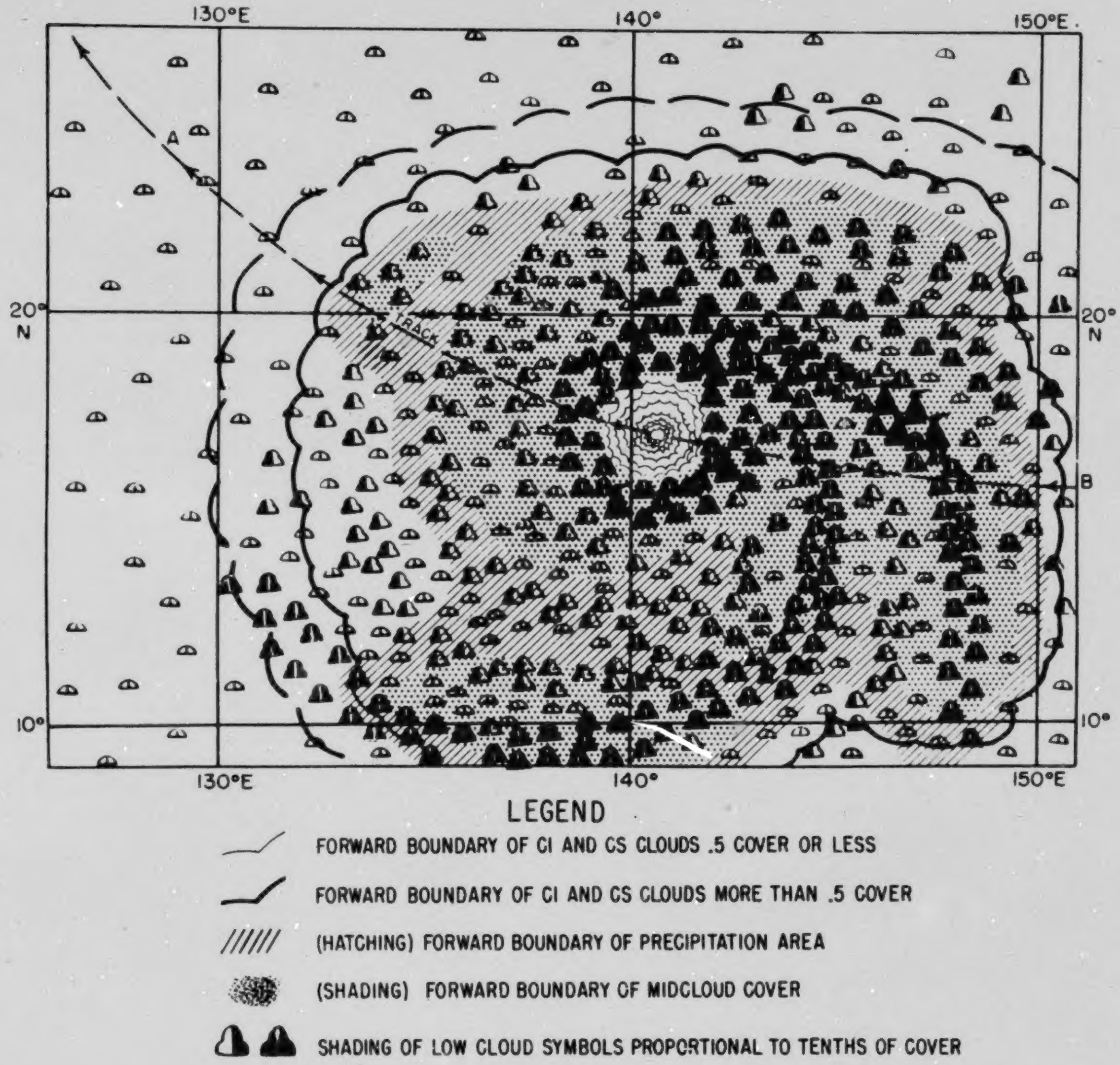
4. Clouds. The cloud patterns of tropical cyclones are also at variance with those of the extratropical cyclones. In mature cyclones almost all the cloud forms are found to be present, to be sure, but by and large the most significant clouds of these storms are the heavy cumulus and cumulonimbus which spiral inward toward the edge of the eye of the storm, becoming generally more massive and closely spaced as they approach the eye. These spiral bands, especially the leading ones, are also referred to as BARS. Cirrus and cirrostratus occupy by far the largest portion of the sky over the tropical cyclone area. In fact, cirrus, becoming more dense, changing to cirrostratus and lowering somewhat, are more often than not the first indicators of the approach of a distant storm or the development of one in the near vicinity. The original appearance of the sky is very similar to that of an approaching warm front. A typical cloud distribution chart for a tropical cyclone is found in figure 12-6.

5. The Eye. The eye of the storm is one of the oldest phenomena known in meteorology. Precipitation ceases abruptly at the boundary of a well-developed eye; the sky partly clears; and the wind subsides to less than 15 knots, and at times there is a dead calm. The sun or the stars become visible. In mature storms, the diameter of the eye averages about 15 miles, but it may attain 40 miles in large typhoons. The eye is not always circular, sometimes it becomes elongated and even diffuse with a double structure appearance. Radar observations indicate that the eye is constantly undergoing transformation and does not stay in a steady state.

6. Precipitation. Very heavy rainfall is generally associated with mature cyclones. However, the methods of measurement are subject to such large errors during high winds that representative figures on the normal amount and distribution of precipitation cannot be said to exist. Precipitation is generally concentrated in the inner core where the slope of the barograph trace is the highest. Amounts of 20 inches are not uncommon. Over the open sea, rainfall is considered of operational interest primarily from the standpoint of its effect on ceiling and visibility. Over land, orographic effects produce concentrations of rainfall, which often results in costly floods. Hurricane force winds forcing moisture-laden tropical air up a steep mountain slope often result in phenomenal rainfall. A fall of 88 inches was recorded during one storm in the Philippines. At the other extreme, as little as a trace has been recorded at a station in Florida, which had winds up to 120 knots during the passage of a hurricane.

7. State of the Sea. One of the first signs of the tropical storm is the swell, which comes in a series of waves with the time interval between crests considerably longer than in waves usually observed in the Tropics. Winds of the storm create waves of the sea which move outward from its center more rapidly than the storm progresses and thus outrun the storm and herald its approach. As the wave moves onward, its height from crest to trough diminishes, its length is reduced, and it becomes a low undulating wave, known as a swell. The size and speed of waves created by the winds depend upon the velocity of the winds and the length of water surface over which the winds blow.

In a tropical cyclone the waves and swell move outward from the storm center at a rate which nearly always exceeds the speed of


130°E
140°
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20° N
10°
A
B
TRACK
130°E
140°
150°E
LEGEND
FORWARD BOUNDARY OF CI AND CS CLOUDS .5 COVER OR LESS
FORWARD BOUNDARY OF CI AND CS CLOUDS MORE THAN .5 COVER
(HATCHING) FORWARD BOUNDARY OF PRECIPITATION AREA
(SHADING) FORWARD BOUNDARY OF MIDCLOUD COVER
SHADING OF LOW CLOUD SYMBOLS PROPORTIONAL TO TENTHS OF COVER

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Figure 12-6.—Typical cloud distribution associated with a tropical cyclone.

progressive movement of the storm. The swell waves continue to move in a straight line through the storm area, whereas the winds turn to the left in the Northern Hemisphere and to the right in the Southern Hemisphere.

The direction of swell waves in the open sea gives some indication of the location of the storm center. When considered in connection with the direction of the wind, the movement of the swell waves is significant.

The period of the swell waves, that is, the time in seconds between the passage of successive swell wave crests, is helpful in determining the intensity of the storm. In the Caribbean Sea and the Gulf of Mexico long period swell waves do not commonly occur except in connection with a tropical storm. The appearance of heavy swell waves with a period of 9 to 15 seconds during the hurricane season in those waters is an indication of the existence of a tropical storm in the direction from which the swell waves come; the longer swell wave periods, 12 to 15 seconds, are almost certain signs of the hurricane. Figure 12-7 shows the direction of wind and swell around the hurricane center.

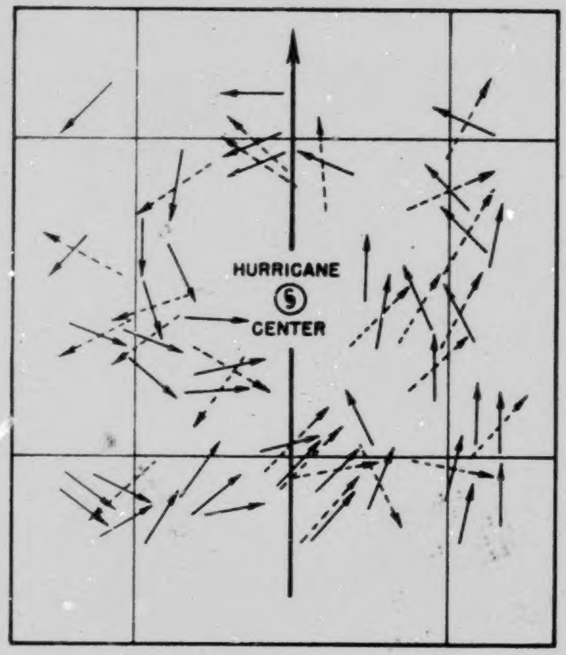

Figure 12-7.—Direction of wind and swell around the hurricane center. Solid arrows show the direction of the wind; dashed arrows show direction of swell. Large arrow shows direction of movement of the storm.

One of the most severe effects of hurricane damage occurs along coastal areas by large ocean waves. The most severe waves can be expected where land partially surrounds bodies of water such as the Bay of Bengal and the Gulf of Mexico. Strong sustained winds in the right-hand semicircle cause a piling up of water along coastal areas as much as 10 feet above normal. Sometimes these tides are referred to as storm tides. An even greater threat is the so-called hurricane wave. The term "hurricane wave" has been applied to the marked rise in the level of the sea near the center of intense tropical cyclones. This rise sometimes amounts to 20 feet or more and affects small isolated islands as readily as continental shores. In partially enclosed seas they may be superimposed on the hurricane and gravitational tides. The hurricane wave may occur as a series of waves, but is usually one huge wave. These waves have produced many of the major hurricane disasters of history. There is usually little warning of their approach. However, they should be anticipated near, and to the right of, the center of intense tropical cyclones.

8. Vertical Characteristics. The vertical structure of a tropical cyclone also differs considerably from the extratropical cyclone. The first difference is that the tropical cyclone is always a warm core low. The storm may build from the top down as well as from the surface upward, and it is for this reason that only the mature model should be considered for comparison. Subsidence occurs in the eye of the storm below about 30,000 feet extending to about 3,000 feet above the surface, accounting for the lack of or sparsity of low and middle clouds.

There is present in a mature storm, considerable horizontal and vertical mixing, extending from near the surface to between 10,000 and 20,000 feet. There is a net horizontal inflow of air at all levels to about 3,000 feet or higher, above which there is a net horizontal outflow of air. This net outflow of air is usually very pronounced in the vicinity of the 200-mb level, and it is for this reason that the 200-mb level is one of the primary analysis tools in the Tropics. The outflow at the 200-mb level is manifested by an anticyclonic flow, unless the storm is unusually severe and penetrates even that level, in which case the flow is cyclonic.

Seasons and Regions of Occurrence

Tropical cyclones may occur during any month of the year with a maximum occurrence from May to November. In general, the summer hemisphere is the region of maximum occurrence for that particular area. Frequency, however, varies from ocean to ocean.

There are eight principal regions of tropical cyclone formation. Five regions are in the Northern Hemisphere; three, in the Southern Hemisphere. The South Atlantic and the eastern South Pacific Oceans are entirely free from tropical cyclones. The southwestern North Pacific on the other hand has by far the greatest number of tropical cyclones developing in it. A detailed breakdown of formation areas, with main tracks and months of most frequent occurrence, is found in table 12-1.

Not all tropical cyclones reach hurricane intensity. Tropical cyclones that reach hurricane intensity are more prevalent in the respective late summer and early fall in both the Northern and Southern Hemispheres. This does not preclude the formation or the intensification of tropical cyclones of any intensity during the other seasons. Tropical cyclones are also much less frequent than the extratropical ones.

SYNOPTIC FLOW PATTERN FAVORABLE FOR DEVELOPMENT

The necessary requirements for tropical storm formation have been researched and discussed to great lengths. It is generally agreed however, that there are at least 3 basic conditions necessary, as follows:

1. Sufficiently large ocean areas with temperatures so high that air lifted from the lowest atmospheric layers and expanded moist adiabatically remains considerably warmer than the surrounding undisturbed atmosphere, at least up to a level of about 40,000 feet.
2. The value of coriolis paramenter larger than a certain minimum value, thus excluding a belt of about 5 to 8 degrees latitude width on both sides of the equator.
3. Weak vertical wind shear in the basic current.

The first requirement limits storm formation to sea or oceanic areas with a surface temperature of approximately 80°F. Statistics verify the second condition in that they show that initial disturbances from which storms later develop may be detected within 5 degrees of latitude of the equator but these disturbances do not intensify into typhoons/hurricanes until they are more than 5 degrees of latitude from the equator.

Additional requirements have been stressed by other researchers. Among these are the presence of a preexisting low level disturbance and a region of upper level divergence or outflow above the surface disturbance. Satellite photographs have verified the preexisting disturbance for many days prior to the development of the tropical storm. Hundreds of tropical disturbances (areas of organized convective activity) occur over the warm tropical oceans each year, yet only a relatively few develop into cyclones of tropical storm or greater intensity. Lack of sufficient upper air data has limited study in the validity of the requirement of upper level divergence.

Some support to the proposed "conditional instability of the second kind" (CISK) as the best physical explanation for storm development exists. This theory stresses the interaction between cumulus convection and large scale motion. A region of cumulonimbus cells in a tropical disturbance warms the atmosphere slightly through the release of latent heat. This heating lowers the surface pressure slightly, thus increasing the low level circulation which in turn increases the low-level convergence because of frictional turning in the boundary layer. This convergence produces more cumulonimbus, causing more latent heating which lowers the surface pressure even more, and so on. In order for this process to be effective, it must occur in a region where the vertical wind shear through the troposphere is very small to allow the released latent heat to be concentrated in a fairly small area. This theory then stresses small vertical wind shear as a primary factor in the development process.

Figure 12-8 shows the average zonal vertical wind shear between 850 mb and 200 mb over the tropics for January, April, August, and October. Due to the predominately zonal easterly or westerly mean flow in the tropics, the mean zonal wind shear, determined by subtracting the mean east-west component at 850 mb from that at 200 mb is similar to the

Table 12-1.—Tropical cyclone data.

		Atlantic Ocean	Pacific Ocean	Indian Ocean
NORTHERN HEMISPHERE	Areas of formation.	1. Cape Verde Islands and westward. Western Caribbean. 2. Gulf of Mexico.	1. Marshall Islands, Caroline Islands, Philippines, and South China Sea. 2. Gulf of Tehuantepec to Revillagigedo Island.	1. Bay of Bengal. 2. Laccadive Islands to Maldive Islands.
	Main tracks.	1. Through West Indies and northward to U. S. or ocean area OR into Central America. 2. West or north.	1. Through or near Philippines and northward toward China, Japan, etc. 2. Northwestward to Lower California or halfway to Hawaiian Islands.	1. Clockwise path into India or Burma. 2. Clockwise path into India or Gulf of Oman.
	Highest frequency months.	Aug., Sept., Oct.	1. July, Aug., Sept., Oct. 2. Sept., Oct.	1. June to Nov., inclusive 2. June, Oct., Nov.
SOUTHERN HEMISPHERE	Areas of formation.	None	Coral Sea and west of Tuamotu Islands.	1. Cocos Islands and westward. 2. Timor Sea.
	Main tracks.	None	Counterclockwise along northeast Australian coast or toward New Zealand. Westward to Coral Sea.	1. Westward, then counterclockwise southward near Madagascar. 2. Counterclockwise along northwest Australian coast.
	Highest frequency months.	None	Jan., Feb., Mar.	1. July, Aug., Sept. 2. July, Aug., Sept.

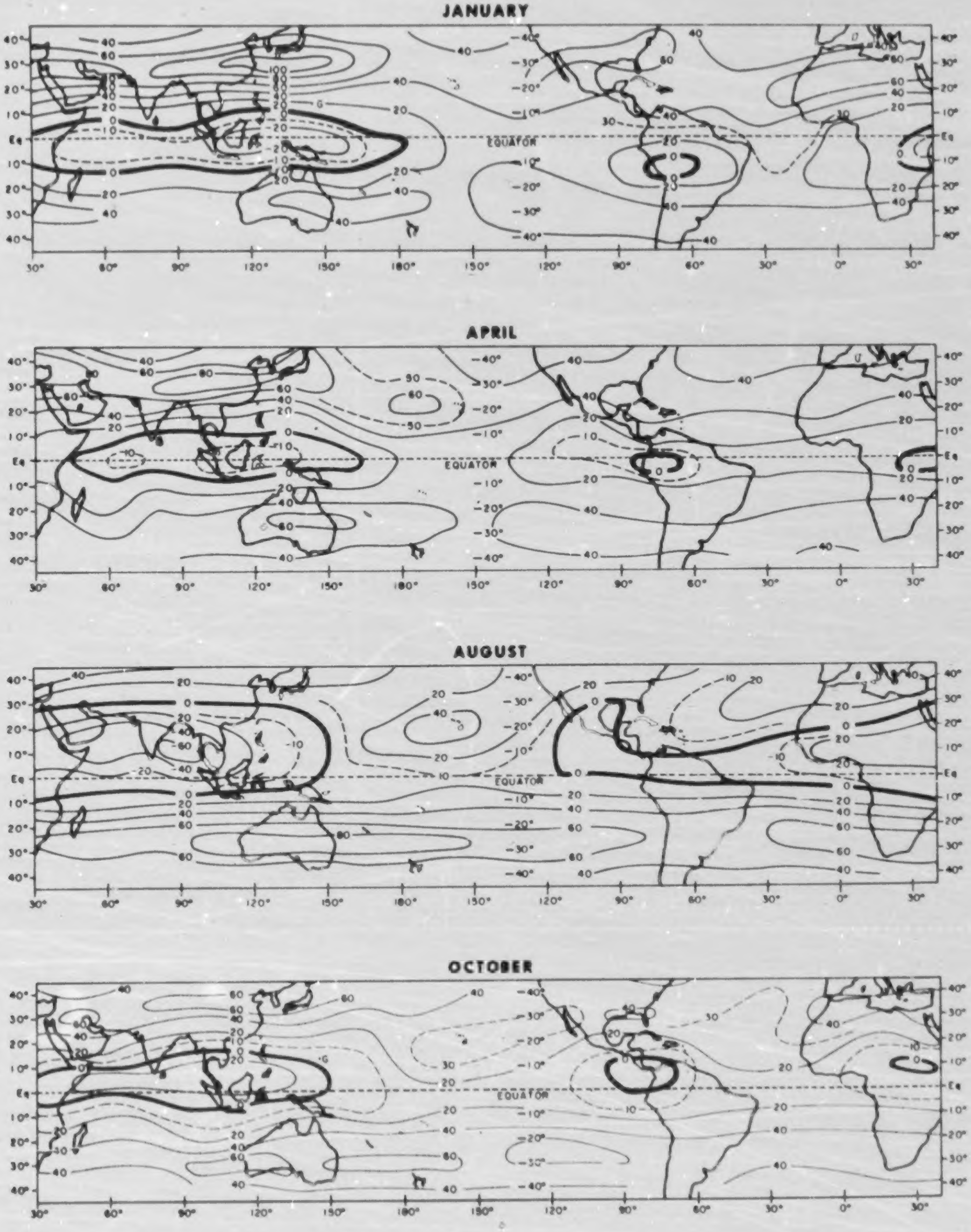
JANUARY
APRIL
AUGUST
OCTOBER
EQUATOR
40°
30°
20°
10°
Eq
10°
20°
30°
40°
30°
60°
90°
120°
150°
180°
150°
120°
90°
60°
30°
0°
30°

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Figure 12-8.—Average zonal vertical wind shear between 850 mb and 200 mb over the tropics.

mean total wind shear obtained from considering both wind components at these levels. The positive values in figure 12-8 indicate that the zonal wind at 200 mb is stronger from the west or weaker from the east.

In general, areas of active storm formation such as the western North Pacific, eastern North Pacific, and North Atlantic (August map) and the South Indian and South Pacific (January map) are regions of small mean zonal wind shear. The data on the chart correlates with the statistical data relating to frequency and time of formation of tropical storms.

In summation there is general agreement on at least 4 of the 5 requirements we have discussed within this section. These are the need for a preexisting disturbance, over a warm (greater than 80°F) ocean area, located more than 5 degrees of latitude from the equator, in a region of small vertical wind shear through the atmosphere.

In spite of this general agreement concerning these requirements for formation, there is still considerable disagreement concerning the synoptic situation in which tropical storms actually form. A number of theories view the development process as progressive intensification of a tropical wave. With the increased observing capabilities provided by satellite pictures during the late 60's none of the storms that developed in the Atlantic were originally associated with a tropical wave. This reduced emphasis on tropical storm formation from tropical waves does not mean that the circulation patterns associated with disturbances which intensify into tropical storms are now completely understood. Recent annual summaries indicate that circulation features in the middle and upper troposphere and their interaction with the lower levels may be more important in this area than in the other storm development regions. It has been discovered that during the heart of the storm season (mid-July to mid-September) most Atlantic tropical storms result from intensification of low-level cyclonic vortices which originate over Africa. These vortices form in the monsoon trough over Africa near the 10,000 foot level and track westward over the Atlantic. In their earlier stages they may or may not show a closed cyclonic circulation at the surface, depending on their intensity. The favored area for intensification of these disturbances is the western North Atlantic as the disturbances move into an area of warmer sea-surface temperatures. Early and late in the Atlantic hurricane season, storms tend to form in the southwestern Caribbean as a result of the eastward extension of the monsoon trough from the Northeast Pacific. In addition to the synoptic mechanisms previously mentioned, storms can be initiated by downward penetration of cold upper lows over subtropical latitudes of the central and western Pacific.

It is evident that a variety of synoptic features are associated with storm formation in the North Atlantic area. In the western North Pacific the low level monsoon trough is the primary synoptic feature associated with tropical storm formation; 85 percent to 90 percent of the storms form in that trough, the upper tropospheric trough being of secondary importance (10 percent to 15 percent of the storms). In all of the other development areas, most of the storms result from the intensification of vortices which form in the low-level monsoon trough.

Estimating the Position of the Storm Center

One of the simplest means of locating the storm center from surface ship synoptic reports is to apply the law of Buys Ballot; that is, when you turn your back to the wind in the Northern Hemisphere, low pressure will lie to your left and slightly to the front. To apply this law you need at least two ship reports from different ships or two reports at different time intervals from one ship. However, from the observations of only one ship, it is not possible to calculate the exact distance of the cyclone center.

An example of locating the tropical storm center from two ship reports is given. (See fig. 12-9.)

From the law of Buys Ballot, with your back to the wind you would expect to find the center of low pressure to the left and slightly to the front (to the west or west-southwest of your position). Swells in the Caribbean Sea do not commonly come from the southwest. The normal condition in that region is an easterly or northeasterly swell, set up by the trades; hence a

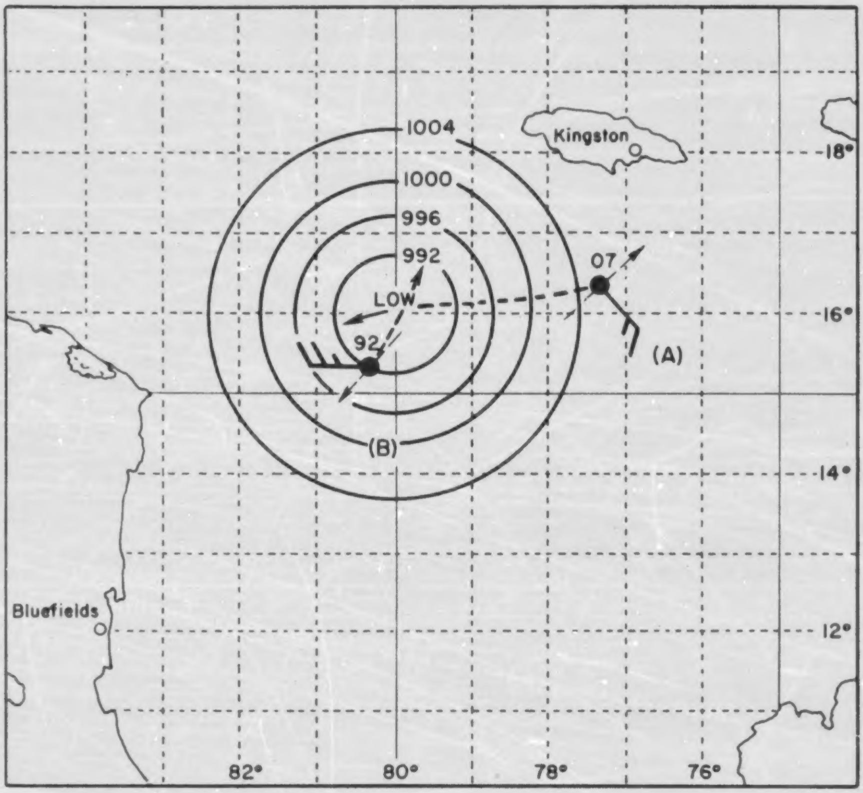

Figure 12-9.—Center of tropical storm located by observations of two ships. Short dashed arrows show direction of waves.

southwesterly swell is strongly indicative of a tropical cyclone. Since we cannot accurately determine the location of the storm center from a single ship report alone, we look for another ship report in the area to use for interpolation of a fix on the storm center.

Obviously ship B is closer to the storm center than ship A. Applying Buys Ballot's law, with your back to the wind on ship B you would expect to find the storm center north or possibly northeast. From these two observations it appears that the storm center was located at approximately 16°N and 80°W. The next step is to tentatively sketch in the isobars as shown in figure 12-9.

The general rule for locating the direction of the center of a tropical storm based on observations of wind direction is subject to certain error when the circulation around the storm is not circular. Often, the wind field around a tropical storm is elliptical as shown in figure 12-10 (A). From a study of the diagram, it is apparent that application of the wind rule may result in large error. Occasionally very elongated wind fields may contain two storm centers. During periods when tropical storm warnings are issued, it is a good idea to copy the fleet facsimile broadcast, the canned map analysis or actual synoptic reports in order to determine the approximate wind field around the storm.

There is an indraft of wind in all tropical storms. The angle of indraft varies with the distance from the storm center; in the outer limits of the wind field it is about 45°; near the

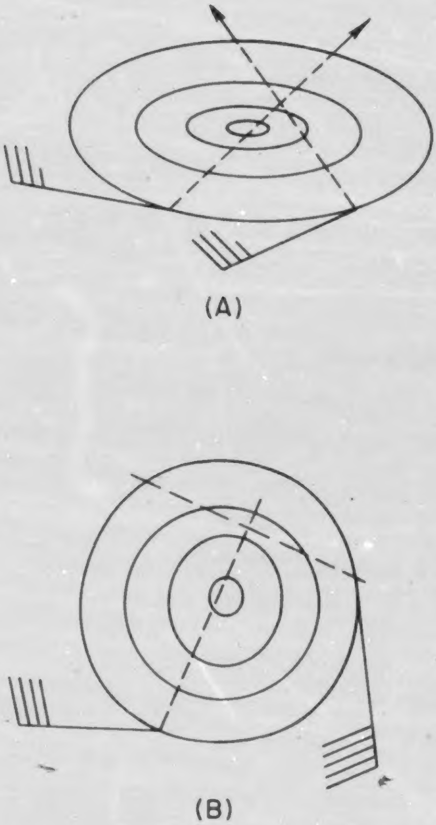

Figure 12-10.—Example of errors in locating direction of storm center. (A) When the wind field is elliptical and the wind rule is used; (B) when the wind rule is applied to fast moving storms.

center the direction of wind movement is more nearly tangent to the isobars.

The angle which the winds blow across the isobars towards the center of a mature tropical storm varies widely in FAST MOVING storms. Directly ahead of the storm, the angle is the least. To the rear of the storm, the angle is greatest. Figure 12-10 (B) shows that the application of the wind rule in the case of fast moving tropical storms will not be as accurate as it is for stationary or slow moving storms.

Satellite Appearance

The cloud systems produced by tropical storms in their early stages of development take on a variety of forms. The shape and appearance of these cloud systems depends on both the type of atmospheric perturbation initiating their convection and the vertical wind shear of the storm's environment. Since the latter is closely related to aspects of the quasi-stationary planetary circulation, such as the subtropical high pressure ridge or the high-level mid-oceanic troughs, storms that develop in different regions of the world appear somewhat different as they evolve. As a tropical storm matures, it modifies its immediate environment to the extent that storms in the later stages of their development look much the same world-wide.

On any one day, numerous synoptic-scale cloud systems appear in the Tropics and subtropics. Only a limited number of these tropical disturbances develop into mature tropical cyclones. A classification system has been developed for these tropical cloud formations. It first seeks to identify those areas of convection which, based on their appearance, seem most likely to develop and, secondly, attempts to classify storms as to their intensity.

Figures 12-11 through 12-19 are satellite pictures of tropical cyclones in various stages of development. The classification ascribed to each is from an earlier method of forecasting tropical cyclone intensity from satellite pictures. The National Environmental Satellite Service (NESS) employs a different system now, but this older system may be used locally as a guide. Reference to this system is contained in AWS TR 240, chapter 8.

TROPICAL ANALYSIS

Tropical analysis consists of complete and detailed analysis of the broad-scale synoptic features as presented on the surface and upper air charts and, furthermore, of more localized and specialized featues as they appear on time sections, space cross sections, low-level and high-level streamline charts, and weather distribution charts. These charts and analyses are the bases for the end-product of the analysis: the forecast.

ANALYSIS OF TIME SECTIONS

Time sections should be kept regularly for all key stations. Their number depends on the size

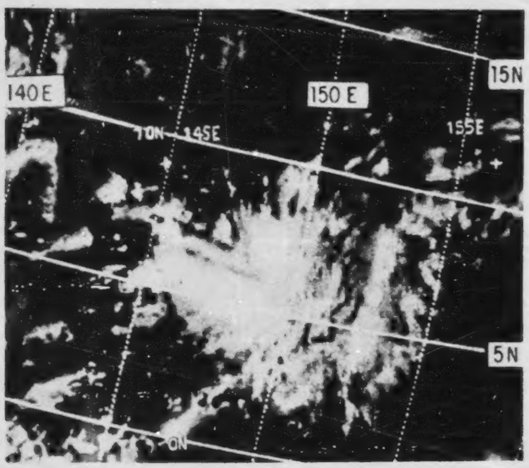
140 E
150 E
15N
10N–145E
155E
5N
0N

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Figure 12-11.—Stage A tropical disturbance, Western Pacific.

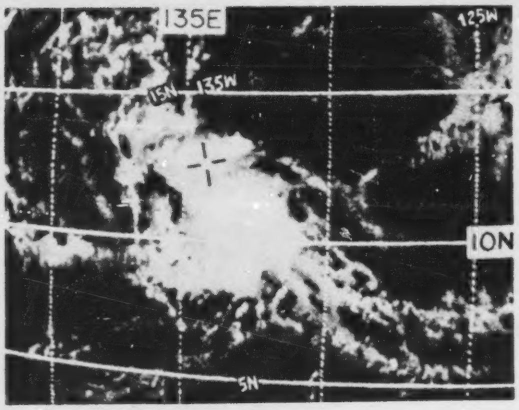
135E
125W
15N 135W
10N
5N

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Figure 12-12.—Stage B tropical disturbance, Mid-Pacific.

of the area to be analyzed and the number of stations available. Up to 20 stations can be profitably plotted and studied on a routine basis. During special situations, additional time sections can be added for periods of a few days.

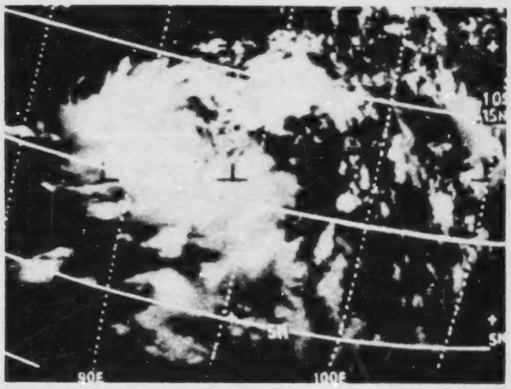
15N
5N
90E
100E

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Figure 12-13.—Stage C tropical disturbance, Western Pacific.

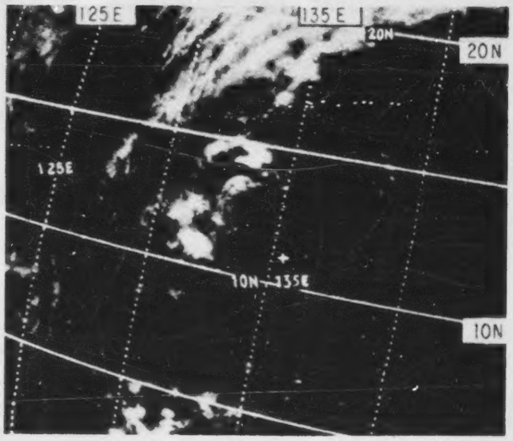
125 E
135 E
20N
20N
125E
10N–135E
10N

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Figure 12-14.—Stage C-minus tropical disturbance, Western Pacific.

The format is important for convenient and rapid handling. Figure 12-20 shows a form for time section analysis.

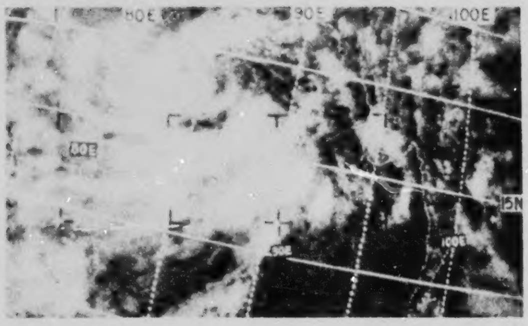

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Figure 12-15.—Stage C-plus tropical disturbance.

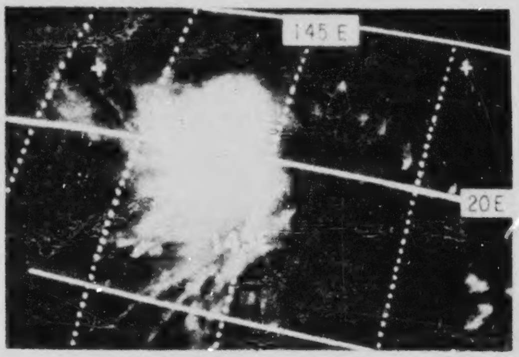

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Figure 12-16.—Stage X, Category 1, Western Pacific.

The vertical coordinate of a time section may be pressure, pressure altitude, or height. It is of advantage to have both a pressure and a height scale, since upper winds are reported at fixed heights, while the significant points on raobs are given in millibars only. Time may be plotted from left to right or vice versa.

Plotting

All upper winds should be plotted on the time sections. Since it is difficult to draw wind arrows quite correctly and since in low latitudes small wind shifts often are important, the observer should write down the coded direction in addition to drawing the vectors as shown in figure 12-21.

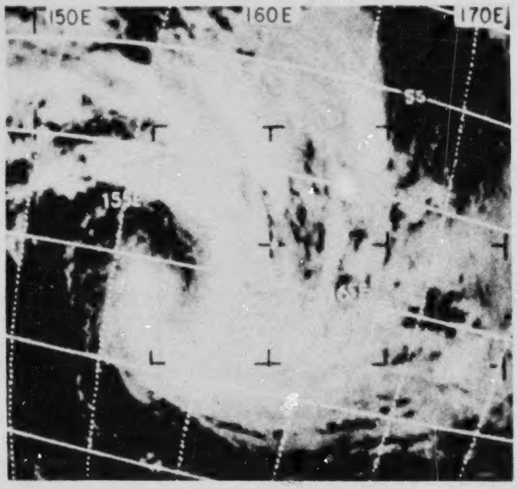

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Figure 12-17.—Stage X, Category 2, South Pacific.

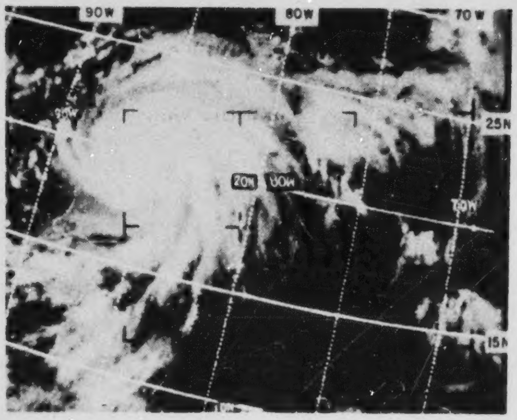

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Figure 12-18.—Stage X, Category 3, Atlantic.

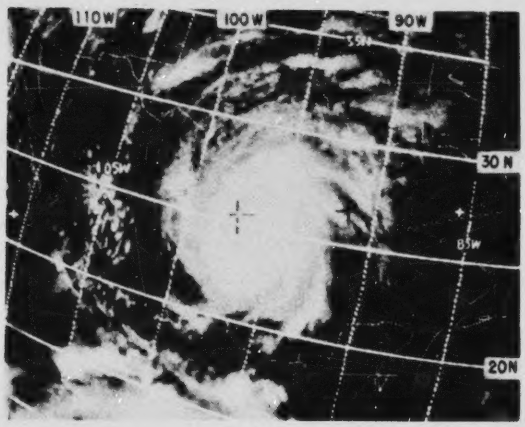

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Figure 12-19.—Stage X, Category 4, Atlantic.

For most purposes, it will suffice to plot winds at 2,000-ft (610 meters) intervals below 20,000 ft (6,096 meters). Higher up, the standard 5,000-ft (1524 meters) interval will suffice.

Analysis

The first object of time section analysis is to detect various errors and unrepresentative values of the reports and make the variations of wind, pressure change, and the like as consistent as possible along the vertical and in time. The second object is to consider surface, upper wind, and raob data together and deduce from them as much as possible about the synoptic situation. Of these the first is by far the easier. Not only will errors usually stand out in an obvious way, but the Aerographer's Mate can also deduce from the time sequence such things as to whether a wind shift in a certain layer is transitory, lasting for only 6 hours, or whether it denotes a longer period change. The sequence of 24-hour height changes of upper pressure surfaces is particularly suited for time section analysis when overlapping 24-hour changes are computed every 12 hours. Considering the normal extent and rate of motion of disturbances in low latitudes, marked upper height falls of 100 feet per 24 hours or more should not be preceded or followed by rises of the same magnitude in the 24-hour interval centered 12 hours before or after, except when accompanied by strong winds and large wind shifts. Otherwise one or more soundings must be suspect. It has already been stated that in such cases the emphasis should be placed on the nighttime data. At reliable stations these changes should be accepted as correct, except (1) when the raob is taken in heavy rain, (2) if a large change is observed, yet there is no previous indication that a large height rise or fall center should arrive, or (3) if the heights rise as much in a second 24-hour interval as they fell in a first without appropriate wind and weather changes.

The first part of the evaluation of time sections, as shown in figure 12-22, is largely qualitative and dependent on an analyst's skill and experience. This is true in even larger measure for the second step, since formal procedures for the integration of time section weather data do not exist.

The following seven semiquantitative steps, however, can be carried out:

1. The principal trough lines and shear lines in the wind field are marked by heavy orange lines and their direction of displacement (especially eastward or westward) is noted with an arrow. These lines give the slope of the disturbances; they will also show the bottom or top of a disturbance, whereas quite frequently a wind shift is mainly confined to either upper or lower troposphere and extends only weakly into the other half.

The distribution of weather as given by the surface reports relative to the time of wind shift shows whether bad weather is concentrated mainly on the forward or rearward side, and how much weather there is. Comparison with time sections where the disturbance had passed previously will (1) furnish the rate of motion, (2) show changes of intensity of winds and wind shifts, (3) reveal changes of weather distribution and intensity with respect to the system. To a lesser degree changes of intensity can also be deduced by variations of the amount of 24-hour surface pressure changes.

2. Isolines are drawn for the 24-hour height changes. The interval chosen should be at least

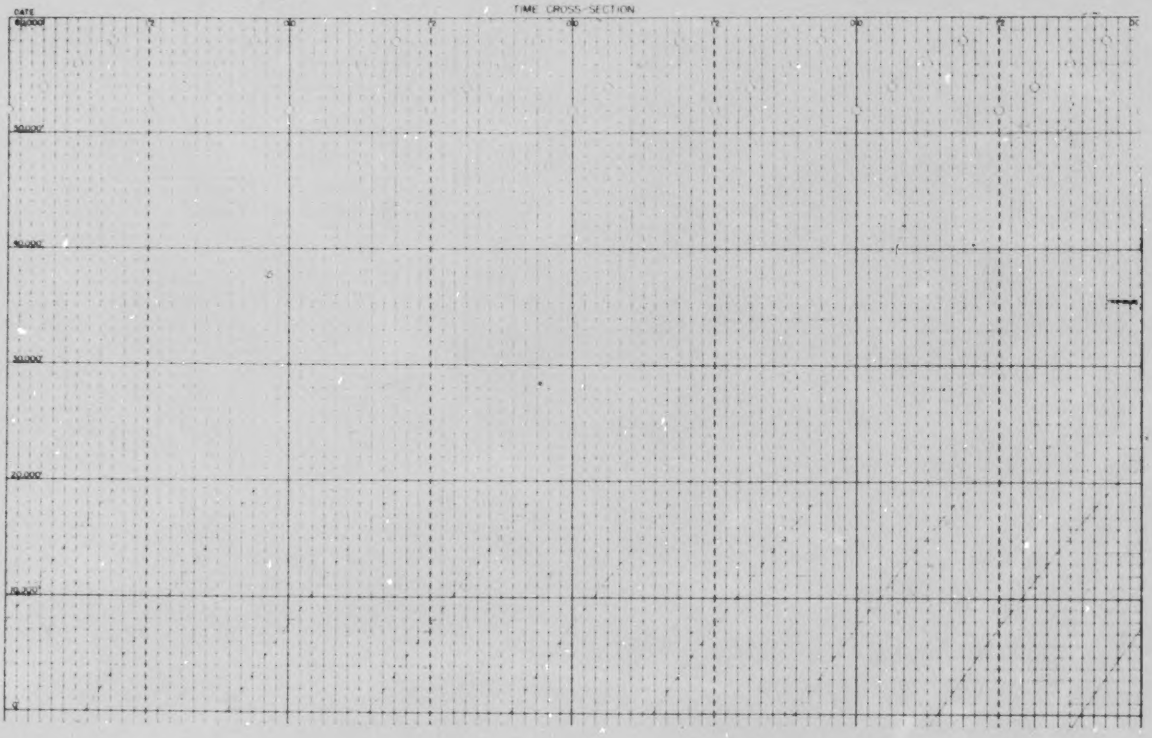

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Figure 12-20.—Form for time section analysis.

100 feet (30 meters), and 50 feet (15 meters) below 400 millibars, since in general changes in the lower atmosphere are smaller than higher up. These changes are examined closely for their relation to the wind shift lines. Trough lines, in particular, must coincide by definition with the instantaneous zero height changes. Normally the 24-hour zero change line should parallel the slope of the wind shift lines and be located not far from them.

The vertical gradient of 24-hour height change is normally correlated with the vertical variation of intensity of moving disturbance. The layer of strongest 24-hour height changes should be the layer of greatest intensity, and during trough or ridge passages the strongest wind shifts should be found there. This statement applies to the vectorial wind shift, not merely the directional change, since the latter can be very large yet insignificant at low speeds. Figure 12-23 shows the computation of the wind shift; its intensity is given by the amount of the vector difference, irrespective of direction.

It is emphasized that the foregoing, while essentially correct in practice, is not always observed and has no necessary foundation in theory. Sometimes there are persistent height falls with little change of wind. This is one indication of deepening of a stationary disturbance, and the situation should be carefully checked for this possibility.

3. The vertical gradient of 24-hour height changes also indicates the areas of cooling and warming, since it indicates whether the isobaric surfaces have moved closer together or farther apart along the vertical. Height changes usually are largest in the high troposphere—both the falls ahead of a trough and the rises to its rear.

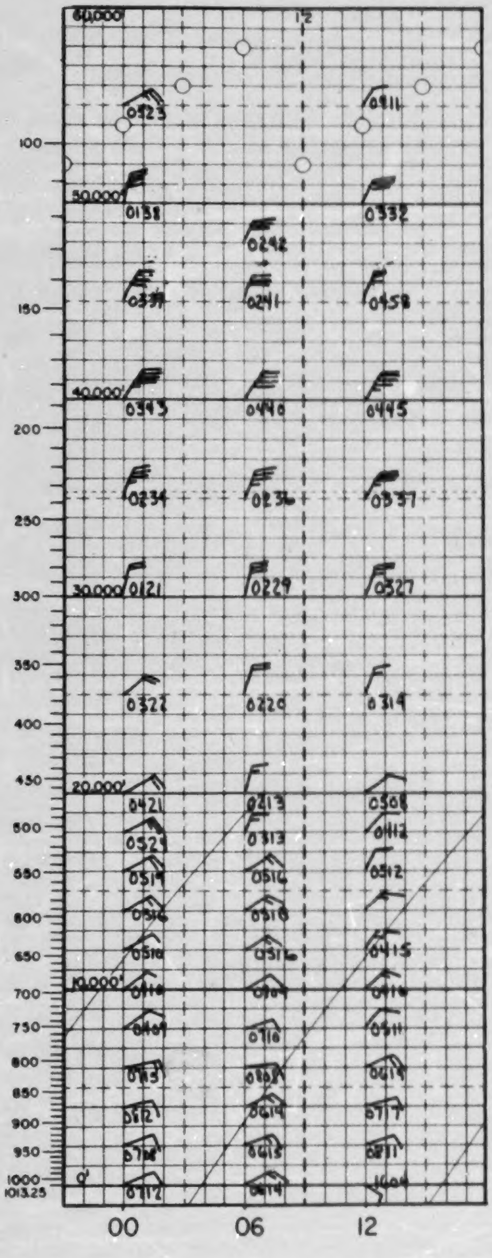

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Figure 12-21.–Time section of upper winds.

This indicates that most troughs have a cold core structure and ridges have a warm core structure. Their intensity increases upward in the troposphere to 200 to 150 millibars; higher up the temperature field reverses and they die out toward the tropopause. This is the main reason why the 200-mb surface is the best choice for high troposphere analysis in the Tropics.

A great deal about the structure of disturbances can be deduced from time section data if the observations are processed carefully. However, the possibilities of analyzing the relations between temperature, pressure, and wind field as given by the time section have not been exhausted. Although the validity of the geostrophic wind relation in the Tropics is questionable, the vertical wind shear appears to give a fair idea as to the distribution of cold and warm air to latitudes 10° and even to 5°. Easterlies decreasing with height denote colder air poleward and warmer air equatorward from the station. Easterlies increasing with height denote the reverse. Northerly winds increasing upward ahead of a trough in the Northern Hemisphere indicate colder air in the trough as do southerly winds increasing upward to the rear of a trough. Winds usually turn counterclockwise with height ahead of such troughs and clockwise to their rear. Thus, the pattern, though generally not the amount of geostrophic thermal advection, agrees with the movement of cold and warm areas. However, actual temperature changes aloft usually are much smaller than those computed from the vertical shear vectors, indicating subsidence in areas of high-level cold advection and ascent in many areas of warm advection. The situation is further complicated by the release of latent heat of condensation in ascent areas. This subject is covered more thoroughly in the upper air analysis section of this chapter.

4. The vertical gradient of height changes can also be related to some extent to the weather. Heaviest activity is likely to occur where lower rises go over into upper falls or where lower falls go over into upper rises. The first of these patterns is found usually with cold core disturbances; it is prominent to the rear of wave troughs in the easterlies. Although the weather may be very intense, the situation usually does not indicate tropical cyclone development. In contrast, lower falls changing to upper rises are the most dangerous of all patterns, since the rain area may develop into a tropical storm. The Aerographer's Mate should at all times be on

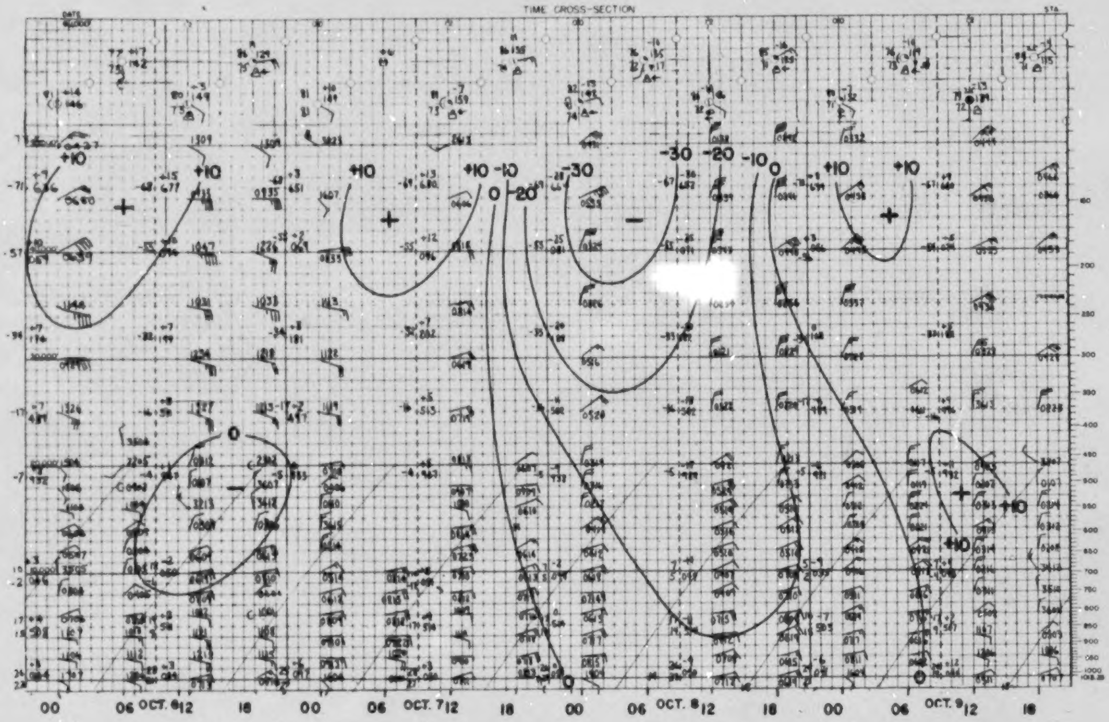

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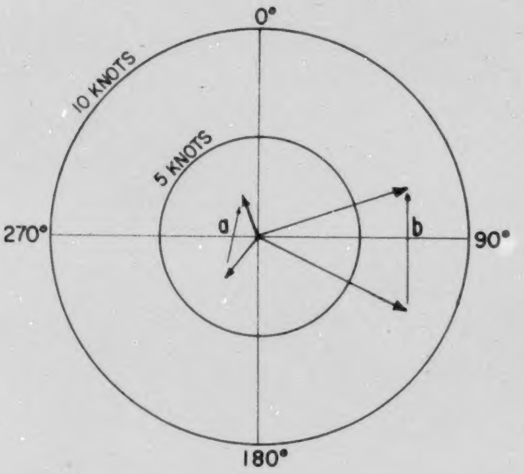
Figure 12-22.—Composite time section (analyzed).

watch for this pattern and also check the passage of a cold core disturbance at successive stations to determine whether the disturbance is losing or gaining cold core characteristics.

5. The depth of the moist layer is indicated as given by the height of the 5 g/kg (grams per kilogram) moisture surface in the rainy season and by the 3 g/kg surface in the dry season. Variations in the depth of the moist layer should be in accord with the observed weather and the deductions made from the preceding steps. Due to the unsatisfactory status of the moisture element on the radiosonde instruments, this correlation will often fail when the moist layer extends above the freezing level. Rather dry air may be indicated even though the balloon is known to pass through thick clouds.

6. A better indication of vertical stability and moisture is bottom, top, and intensity of stable layers, especially the trade inversion. The potential temperature difference between top and bottom, and the inversion thickness in millibars, are the best measures of the inversion strength. The lid against penetration and destruction of the inversion will be the larger, the greater the increase of potential temperature through the inversion and the thicker the inversion layer. The inversion analysis establishes the current average cloud top height and gives an indication to what extent an approaching high-level disturbance may be able to raise and weaken the inversion and raise the cloud tops.

7. Base and top of equatorial and polar westerlies may be marked as a final step, mainly to indicate clearly the depth of any layers with westerly winds and the time changes. This is useful in determining and predicting the direction of motion of disturbance. These move eastward when a deep layer of westerlies is present, westward in deep easterlies. A change in the thickness of a layer of easterlies or westerlies may indicate a reversal of the direction of



(a) LARGE DIRECTIONAL CHANGE WITH LOW SPEEDS.
(b) SMALLER DIRECTIONAL CHANGE WITH HIGHER SPEEDS

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Figure 12-23.—Illustration of a wind change.

motion, if the change is representative for a large area. The temporary appearance of deep westerlies at one station during passage of a cold core low to the north cannot be interpreted in this way.

This completes the time section analysis routine as far as it can be described. If the Aerographer's Mate operates in an area with a reasonable station network, he will, after going through his time sections, have acquired a fairly definite knowledge of how the charts should be drawn and where the most dangerous areas are located.

LOW-LEVEL STREAMLINES

The level to be chosen for the low-level streamline analysis may vary from 2,000 to 5,000 feet, depending on several factors. In general, the wind frequency is fairly equal in this layer, but since the cloud bases average about 2,000 feet, balloons are lost above this level when low cloudiness is great. Ships' surface wind can be used to supplement 2,000-ft but not the 5,000-ft winds. In the equatorial trough zone, the sharp shear found between opposing currents near the ground often diminish markedly with height between 2,000 and 5,000 feet. On the other hand, the lower level is much more subject to orographic influences, and in the trade region, where disturbances often damp out downward to the ground, the 2,000-ft wind fluctuations may be small and noninformative. Thus, the choice of level must be based on the local and synoptic peculiarities of the area to be analyzed.

Plotting

Plotting consists in entering wind arrows and barbs, and wind direction and speed written in code form, in parenthesis, alongside the barb. Some stations may vary the last entry. The use of a protractor to plot information for key stations is recommended.

A 12-hour continuity of the streamline charts will usually suffice; the data from the off-periods are used to supplement the ontime reports for continuity and when the latter are missing. The ontime periods should be chosen nearest the raob observation times, but this depends on local scheduling of observations and may not always be feasible.

Analysis

The objective of wind analysis is to construct a continuous representation of the wind field from the observation of the two-dimensional horizontal vectors on the surface in question. The complete analysis of the wind field also provides a means of obtaining other subsidiary fields such as the divergence, convergence, and vorticity of the wind.

There are two basic methods of streamline analysis in use: the discontinuous or qualitative method and the streamline-isotach method. If reports are sparse the discontinuous or qualitative method is generally employed. The most complete analysis is made with the streamline-isotach method and this is the more recommended procedure.

DISCONTINUOUS OR QUALITATIVE STREAMLINE ANALYSIS.—This method involves a single set of lines drawn tangential to the wind direction and spaced in proportion to the wind speed. This means that in areas of speed convergence some lines must be dropped from the field and in areas of speed divergence some must be added to this field; thus the streamlines are discontinuous. It is easily seen that it is impossible to represent completely the true wind field in this manner. Figure 12-24 illustrates this method of wind analysis.

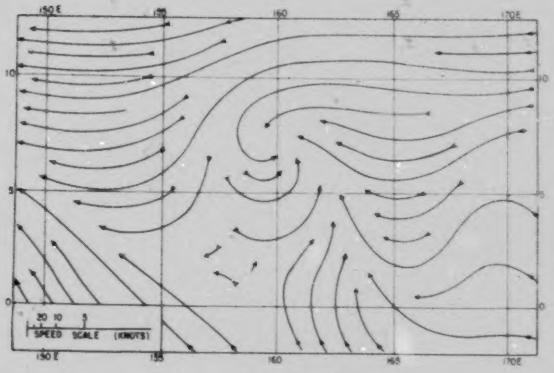

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Figure 12-24.—Discontinuous or qualitative streamline analysis.

STREAMLINE-ISOTACH ANALYSIS.—This method consists of two sets of lines: streamlines representing wind direction and isotachs representing the wind speed, labeled in knots. The two sets of lines give a continuous representation of the wind field from which the forecaster can normally determine wind direction and wind speed at any point on the chart.

In the streamline-isotach method of analysis familiarity with circulation patterns is a necessity. The following are definitions of some of these circulation patterns that will be encountered:

1. Asymptotes. These are streamlines in the wind field away from which neighboring streamlines diverge (positive asymptotes) or toward which they converge (negative asymptotes). Asymptotes may or may not represent lines of true horizontal mass divergence or convergence depending upon the distribution of wind speed in the area. Therefore it is better to use the terms asymptotes of difluence and confluence rather than divergence or convergence. Typical examples of asymptotes are shown in figure 12-25.

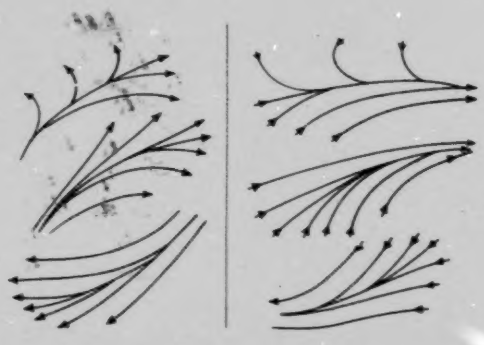

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Figure 12-25.—Streamline asymptotes of difluence (divergence) left; of confluence (convergence) right.

2. Waves. These are perturbations in the streamlines analogous to the wavelike arrangements of troughs and ridges in isobaric patterns. Waves that do not extend across the entire width of the current in which they are imbedded are called damped waves. In this case the streamlines on one or both sides of the current have smaller amplitude than those in which the wave is more pronounced. Figure 12-26 illustrates a damped wave in the streamlines.

3. Singular Points. These are points into which more than one streamline can be drawn or about which streamlines form a closed curve.

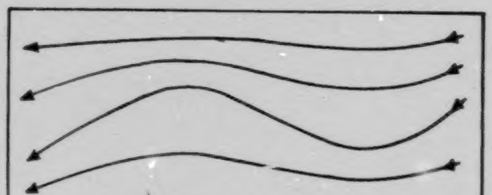

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Figure 12-26.—A damped wave in the streamlines.

The wind speed is zero at singular points and the speed immediately adjacent to the point is relatively light. There are three classes of singular points: cusps, vortices, and neutral points. These are described as follows:

a. Cusps. An intermediate pattern in the transition between a wave and a vortex. They are relatively unimportant in synoptic wind-analysis since they are short-lived and there is normally insufficient data to determine their presence. Figure 12-27 illustrates two variations of this class of singular points.

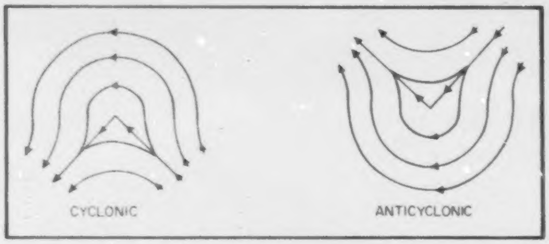

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Figure 12-27.—Cusps.

b. Vortices. These are cyclonic or anticyclonic circulation centers and anticyclonic (cyclonic) outdrafts or anticyclonic (cyclonic) indrafts. Figure 12-28 illustrates four of the commonly observed vortices.

At low levels, anticyclonic outdrafts and cyclonic indrafts are frequently found. Data is usually too sparse at upper levels to determine the outdraft or indraft characteristics of vortices. Therefore, many upper vortices are drawn as pure cyclones or anticyclones for lack of more detailed information.

c. Neutral Points. These are points at which two asymptotes, one of directional confluence (convergence) and one of directional difluence (divergence) come together. They are analogous to cols in that they represent a saddle between two areas of anticyclonic flow and two areas of cyclonic flow. Neutral points in the streamlines are shown in figure 12-29.

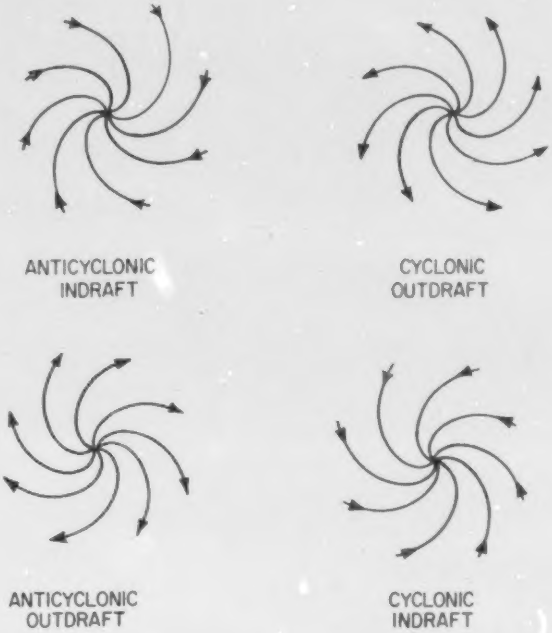

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Figure 12-28.—Vortices in the streamlines (Northern Hemisphere).

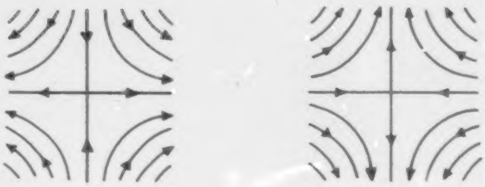

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Figure 12-29.—Neutral points in the streamlines.

PROCEDURE.—There are two basic methods of performing streamline analysis: the isogon method and the direct method. The isogon method employs the use of an intermediate set of lines called isogons. An isogon is a line joining points in a plane which have the same wind direction. The isogons are drawn according to the reported wind directions with interpolation being carried out continuously as in scalar analysis. Then short lines are ruled on each isogon corresponding to the wind direction it represents. Streamlines are drawn utilizing the original wind direction and interpolated isogons. This method, although the most accurate, is too

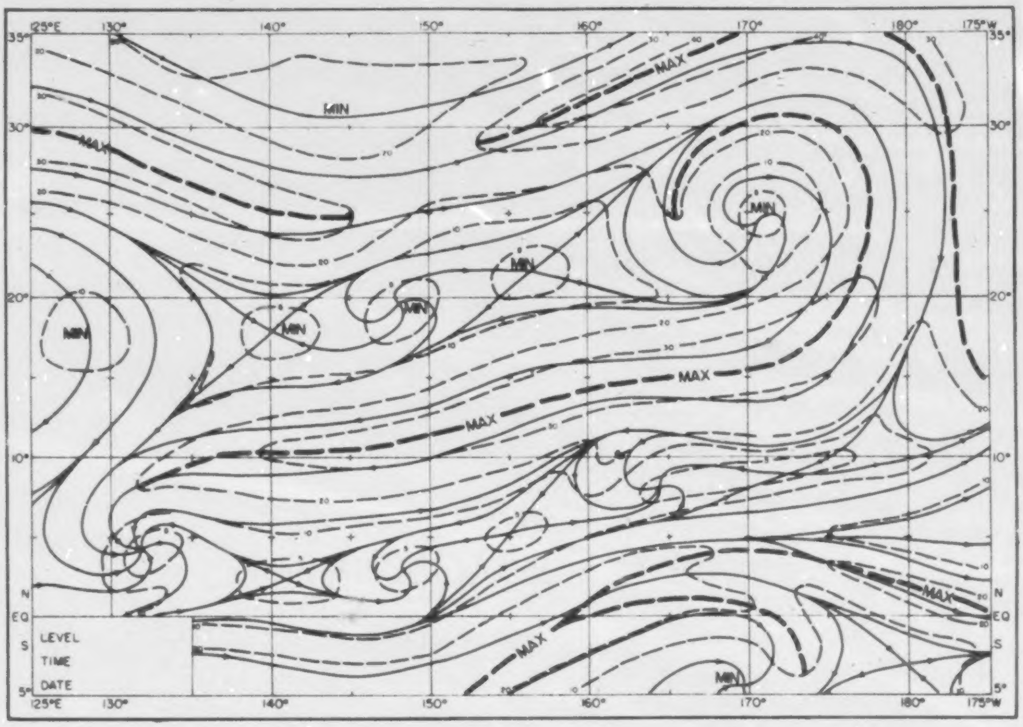

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Figure 12-30.—Streamline-isotach analysis over a large ocean area.

time consuming to be in general use. However, it is recommended that beginners use this method to achieve familiarity and experience with streamline analysis.

In the direct method of streamline analysis interpolations among the winds are carried out by eye and the streamlines are drawn directly using a trial and error approach. The accuracy achieved depends to a great extent upon the skill of the analyst.

Isotach analysis is performed after the preliminary streamline analysis is completed. Isotach patterns will resemble those found in simple scalar analysis, i.e., there are centers of maximum and minimum values and saddles or cols in the wind field.

Figure 12-30 shows a completed streamline-isotach analysis.

There are a number of guidelines as well as basic rules concerning streamline analysis that the analyst should follow to achieve the most complete and accurate product. A comprehensive listing of these as well as detailed information concerning streamline analysis is contained in AWS Technical Report 240, Forecaster's Guide to Tropical Meteorology.

SURFACE CHART

The low-level streamlines, especially when at 2,000 feet, determine the character of the surface isobars to within about 5° of the Equator; in the center of the equatorial zone the analyst must rely on the pressure reports alone. In fixing the course of the isobars, the low-level streamlines are thus an invaluable aid in sorting out good and bad pressures, especially in the trades. Nevertheless, since they are concerned exclusively with the wind field, they do not convey information about the field of mass and its time changes, which is essential for many prediction problems.

Pressure change charts are considered the most important aspect of the pressure analysis. However, in order to draw isallobars, it is necessary to draw isobars first and compute pressure changes in the areas where stations are widely spaced. If this is not done, the isallobaric centers will inevitably be put near the observing stations, and their paths will follow any lines of stations that may exist, such as the Antilles. You also need the isobars for the differential analysis of the upper isobaric surfaces. Although the value of the isobars to the forecaster has often been questioned, they remain a necessary tool. In the following discussion, consider these three steps: Surface isobaric analysis, surface 24-hour pressure change analysis, and weather distribution analysis from surface reports.

Surface Isobaric Analysis

When sea level pressures have been plotted with the careful selection routine described earlier, the problem of ship pressures still remains. To date, the most successful remedy has been to follow the path of individual ships and even plot their reports in time series. This will tell quickly whether reliance can be placed on the pressures from a certain vessel or whether a constant correction should be applied. Isolated reports from ships suddenly appearing in the middle of the area of analysis should receive least weight. As a further aid, sea level pressures reported by low-flying reconnaissance aircraft should be plotted on the surface map, together with the winds at flight level. In doing so, however, the diurnal pressure variation during the flight time must be considered, though synoptic pressure changes over a time interval over 3 hours before to 3 hours after map time may normally be disregarded safely.

However, 3-hour pressure changes south of 15°N over the western Pacific may amount to as much as 2 to 3 millibars or more. This can be taken into consideration by knowing the average 3-hour tendencies for a particular area or station. Surface pressures plotted from dropsondes should also be plotted in the surface chart.

The diurnal variation is particularly troublesome when charts extending over 90° long, or more are drawn at 6-hourly intervals and then compared in sequence. Troughs and ridges of the semidiurnal pressure wave are 90° long. apart. Suppose the normal pressure gradient in a tropical area is 3 millibars per 10° lat. and the isobars run straight east-west at a given map time. If a ridge of the semidiurnal wave reaches the western end of a 90° long. interval 6 hours later and a trough reaches the eastern end, the isobars at that time will trend equatorward by no less than 10° lat. from eastern to western end of the map at that time, if the pressure difference from crest to trough is about 3 millibars. (See fig. 12-31.) Further, the trough area will feature cyclonic isobars in a huge arc—this is especially noticeable in the western Pacific. As seen from figure 12-32 the 1,010-mb isobar is displaced by approximately 1,500 miles in 6 hours. Evidently, 6-hourly maps cannot be compared under such circumstances.

A remedy for this difficulty has been suggested. At any point of the map the normal departure of pressure from the mean can be computed for the normal map times of the day. If this is done for the whole area of analysis and for each month of the year, the correction that should be added or subtracted from the current sea level pressure wave can be determined. Although this wave is not quite constant from day to day, charts drawn with such corrections will be fairly comparable.

It is essential to draw for 2-mb intervals in the trades and for 1-mb intervals close to the Equator. In addition, the 1-mb interval should be used whenever the isobars are so widely spaced that even 2-mb isobars do not reveal the important features of the pressure field. It is thus readily seen how important it is to screen the pressure reports carefully. A few bad pressures can distort the isobaric pattern in a grotesque way. All pressures of normally reliable stations should be drawn to quite strictly and disregarded or modified only when excellent reasons exist. Too often one approaches a chart with an initial idea as to how it should look. It is much better to let the data guide the pencil.

For application to differential analysis, the finished isobars must be relabeled in terms of the height of the 1,000-mb surface. This can be accomplished with the standard method of 7 1/2 millibars equal 200 feet (60 meters), starting

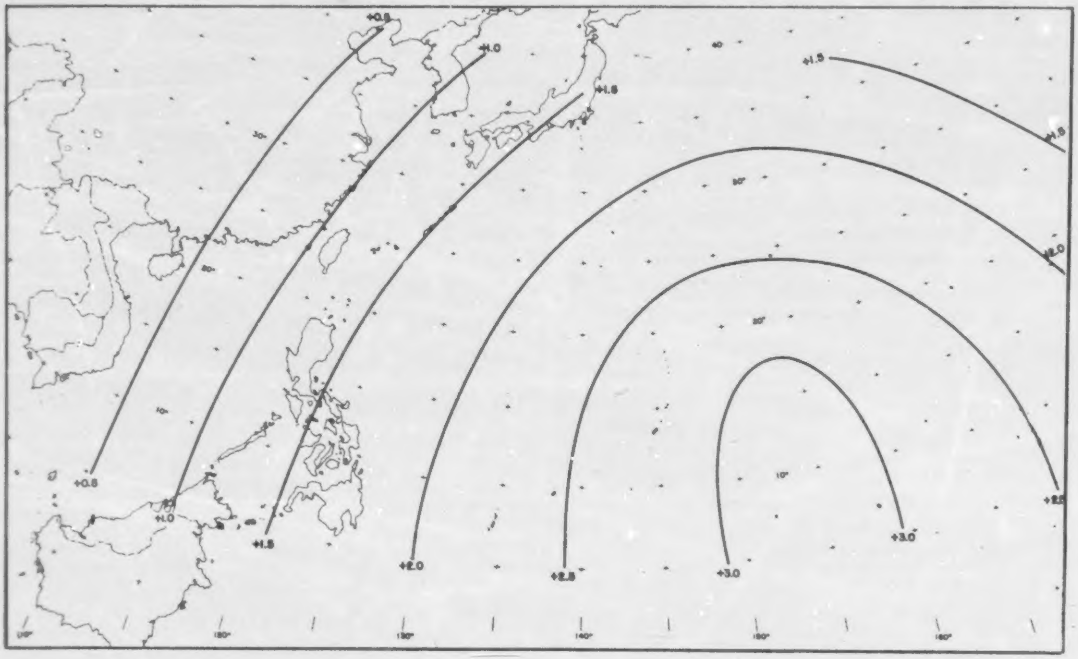

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Figure 12-31.—Six-hour diurnal pressure change from 0000Z to 0600Z.

from 1,000 millibars; when greater accuracy is desired, use table 12-2. Table 12-2 has been calculated, using the normal temperatures prevailing near the surface over the tropical oceans. The isobars are labeled directly in units of the 1,000-mb height.

24-Hour Pressure Changes

The 24-hour changes observed at all reporting stations are computed and written just above the pressure values. This should be a step of the normal plotting routine. It will help to circle these changes so that they will stand out better among the mass of information on the surface map. Over the ocean area, the procedure is to tabulate the pressure daily at all 5° latitude-longitude intersections, to compute differences, and to plot these. Isallobars are then drawn at intervals of 0.5 to 1.0 millibar, depending on the magnitude of the changes.

Alternately, the changes may be obtained by graphical subtraction of the current chart from the previous one. This, however, will often prove less satisfactory, since the isobars tend to run fairly parallel and intersect usually at very small angles.

The main objective of the analysis is to track rise and fall centers and note their changes in intensity. In addition, it furnishes a check on the ocean isobaric analysis; if large regions of intense falls suddenly appear without continuity, the Aerographer's Mate will do well to reexamine his work.

In general, the isallobaric analysis will yield good continuity. It can even be extended to such areas where pressure values are quire irregular due to lack of precise knowledge of station altitudes.

Usually, one looks for concentrated fall or rise centers covering limited areas only to locate intense disturbances. Wide areas of fairly uniform falls are considered harmless.

Occasionally falls of 1 to 2 millibars over most if not all, of the region of analysis will

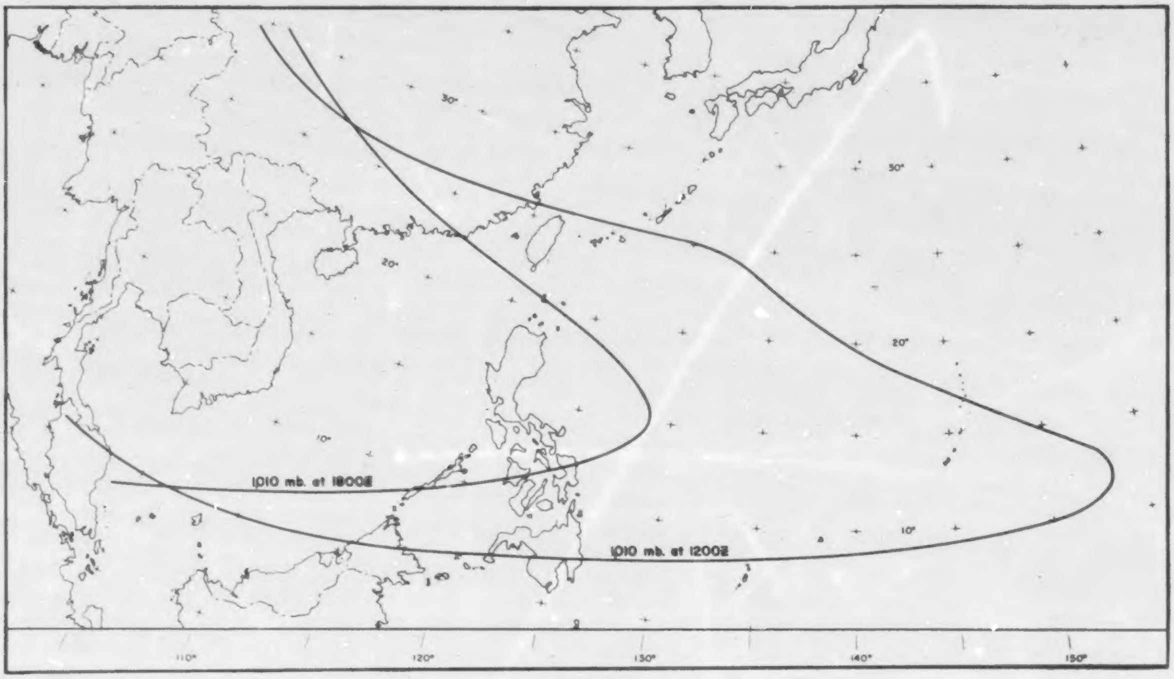

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Figure 12-32.—Sample normal displacement of 1,010-mb isobar between standard map times.

appear suddenly on 1 day, to be canceled by corresponding rises within 24 to 48 hours. Such pressure waves on a huge scale are not understood at present, but for forecasting they appear to be without significance. When such widespread falls occur, a synoptic fall area will appear as a center of still larger falls, but a synoptic rise area may take the form of an area of small falls embedded in surroundings with larger falls.

Subsequent to the analysis, the isallobaric centers should be related to the low-level streamlines and the weather pattern, later they should also be compared with the distribution of wind and temperature aloft. In particular, fall centers containing much bad weather and associated with warming in the troposphere and anticyclonic circulation at 200 millibars must be regarded as potentially dangerous, especially when there is a trend toward an increasing association of these features. Rise areas associated with bad weather generally do not indicate immediate deepening even though the intensity of the convection may be severe. The forecaster should be on the lookout for rises surrounding an area of weak falls or even no changes. Especially if the rises are intensifying but not spreading in area, sudden intensification of the falls is apt to follow.

Weather Distribution

The weather distribution map is a graphical representation of the clouds and weather over an area of operation. In high latitudes, most forecasters would be reluctant to spend the time necessary for analyzing such a chart because the surface synoptic chart, with the clouds and weather plotted at each station together with the identification of air masses and fronts, is adequate for most purposes. For operations in the Tropics, however, the weather distribution map is almost indispensable for both forecasting

Table 12-2.—Sea level pressure versus 1,000-mb height.

Sea level pressure (mb)	1,000-mb height	
	(tens of ft)	(meters)
1,000	0	0
02	6	18
04	12	37
06	17	52
08	23	70
10	29	88
12	34	104
14	40	122
16	46	140
18	51	155
20	57	174
22	63	192
24	69	210
26	74	226
28	80	244
30	85	259

and flight briefing. In instances where precise and detailed analysis of weather over particular terminals or areas is required, it is often necessary to follow the continuity of the weather and cloud patterns in the same fashion as the continuity of low-pressure centers and fronts. As a briefing chart, the weather distribution map is perhaps the most practical means of presenting the weather to nonmeteorologists. Finally, the art of forecasting the weather depends upon the ability to correlate changes in the wind and pressure patterns with parallel changes in the cloud and precipitation patterns.

Several methods of drawing a weather distribution map have been developed. All involve the use of the standard synoptic cloud symbols to designate the cloud types. The deviations are mostly confined to the representation and analysis of the amounts of clouds. The choice of method can be determined locally. The data included on the plotted charts should be as complete and extensive as possible. Besides the regular land and ship synoptic reports, other reports entered on the map are reconnaissance, off-hour ship, pirep, and aircraft in-flight reports, each appropriately designated as to type and time of report. Past weather should also be entered on the weather distribution charts, for it will aid in correlating the movement of weather systems, and may fill blank spots on the charts on occasion.

ANALYSIS WEATHER DISTRIBUTION.—If weather distribution maps form a part of the standard daily analysis in the weather office, the task of drawing the map is greatly simplified by the fact that the major weather systems of the Tropics have continuity from map to map. From day to day, large-scale systems moving through the area of operation may show the same characteristic intensity, as judged by the amount, depth, and arrangement of the clouds and the precipitation patterns. On the other hand, they may increase in intensity or die away. However, during the explosive deepening of some typhoons and hurricanes, the changes are sufficiently slow to be followed easily on the map sequences. In some regions, systems may form aloft and remain stationary for many days, slowly increasing in intensity from day to day. Under these circumstances, the whole process of deterioration in the weather can be followed in detail and forecast by simple extrapolation.

The following discussion of weather distribution analysis is limited to times when weather distribution charts are not part of the normal routine of the weather office and the Aerographer's Mate, as analyst, must start from scratch.

1. If the analyst has not already become familiar with the climatology of the area, he should consult the best available information on the cloud and weather characteristics of the

season and place. If the equatorial trough lies near or across the area, he should be on the alert to discover in the reports evidence of "equatorial fronts"–lines of cumulus congestus or cumulonimbus that simulate the cold fronts of high latitudes. If he is working in a trade wind area, he should be prepared, before starting the analysis, to find characteristic distribution and heights of cumulus and stratocumulus. In other words, his climatological knowledge should help him form a mental picture of the normal weather distribution map of the region. If he finds radical departures from this pattern, he will know that he must pay particular attention to the anomalous features, making every effort to delineate them properly on the map.

2. Examine the aircraft reports, first priority being given to any reconnaissance data. If, as is customary in some regions, transient aircrews have submitted pictorial cross sections of the weather along the air routes to the station, these should also be examined at this time. Particular attention should be given to plain language description changes of cloud form, as these usually indicate the extent and orientation of major cloud systems. The analyst should outline these features lightly. After the aircraft reports have been thoroughly examined, analyze the remaining reports in approximately the same fashion.

3. Outline the areas of middle clouds. At first, all areas of middle clouds should be outlined by following the reports rather mechanically and making reasonable interpolations between reports. Then, the resulting picture should be examined with an eye to the reports of cumuliform clouds. If cumulonimbus are absent, any mid cloud is probably independent. Some attempt should be made to distinguish an organized shape to the system. Independent systems will usually be bandlike or sickle shaped with frayed edges of patchy altocumulus.

All middle clouds will not be independent. If orographic cumulus or cumulonimbus are widespread, fairly large patches of middle clouds may be reported in their neighborhood. However, these patches will be detached from one another and will rarely form a very extensive alto-system with a definite shape. If reports of cumuliform clouds indicate that the middle cloud in any part of the region is dependent, the analyst should keep the area covered by his drawing of the cloud to the minimum compatible with the reports and the topography; at the same time he should show clearly the connection between the mid cloud and the convective pillars. In passing, it should be emphasized that over the open sea, lines of cumulus congestus or cumulonimbus are often associated with alto-systems, but this does not necessarily mean that the mid cloud is dependent; on the contrary, the most likely situation, during the late stages of deepening of an upper-level cyclone, is for one or more asymptotes of convergence, with accompanying cumulus or cumulonimbus lines, to develop under a preexisting deep alto-system and merge with it aloft.

4. Delineate the high clouds. Follow the same principles as were used in drawing the mid clouds. Usually there is little difficulty in distinguishing between dependent and independent cirrus. Almost all independent alto-systems will be accompanied by cirrus or cirrostratus sheets, either separate and at a much higher level, or fused with it in the areas of precipitation. Usually the cirrus sheets cover a wider area than the alto-system, and often cirrus in broad bands cover a great area on either side of the alto-system, the bands being oriented parallel to the main axis of the alto-system.

The delineation of cirrostratus sheets of independent formation can be of great assistance in tracing the genesis and development of an upper-level cyclonic system. At early stages in the development of such systems, very little mid cloud may be present, and the circulation may be evident only in the layers above 30,000 feet (9,144 meters). On the weather distribution map, however, this early stage of development is often accompanied by extensive sheets of cirrostratus. A series of maps which clearly shows the development of the cirrostratus sheets is often the best way to follow the gradual intensification of the system.

5. By the time he has analyzed the distribution of the mid and high clouds, the forecaster should have studied most of the lower cloud reports and should have formed definite opinions about their distribution. Low clouds of orographic origin should be obvious by this

time. The more definite frontlike lines of cumulus and cumulonimbus should be clearly indicated. Often, very extensive areas may be covered by more or less uniform distributions of cumulonimbus. Delineate these areas as sharply as possible.

6. Next, note all precipitation and special phenomena. The analysis should then be complete.

Although various combinations of cloud types and amounts will occur, the following three patterns tend to be most prominent:

1. A large amount of low clouds, cumulus congestus and cumulonimbus with mid and high clouds and reports of current or past showers or rain. This pattern, of course, is typical of disturbed areas.
2. Few or no mid or high clouds, either with few low clouds, or with large amounts of cumulus humilis (also called fair-weather cumulus) or stratocumulus. This combination indicates suppressed convection.
3. Low clouds near average, occasional cumulus congestus, and mid and high clouds in varying amounts. In such a transition pattern, it is most important to study the upper clouds, since these are independently formed in such cases rather than derived from tall cumuliform types. Thickening cloud cover aloft, especially altostratus overcast, denotes an approaching, forming disturbance.

CORRELATION OF WIND AND WEATHER.–The correlation of wind and weather is still in the infant stage of development, and rules evolved do not always have universal application. Local peculiarities tend to necessitate some modification and in many instances require rules of their own which do not lend to generalization. However, there are some general rules available, and, surely, future experience with wind and weather correlation will yield many more.

In as much as the vast majority of clouds are formed by upward motion of air and vertical convergence of air can be deduced from the horizontal components of direction and speed of the wind field taken over large areas, we find that many of the weather patterns in general, and specifically in the Tropics, have a direct relationship to streamline patterns at all levels.

Complete correlation of wind and weather requires analysis of the wind field at several levels. Data limitations and time limitations usually allow a correlation for only two levels: the low-level streamlines between 2,000 to 5,000 feet (610 to 1,524 meters) and the streamlines of the 200-mb or 40,000-ft (12,191 meters) level. Following are some correlation rules.

In regions of moderate to strong divergence at the low-level streamline level, the predominant cloud is usually cumulus humilis with less than one-half coverage.

When great vertical shear accompanies moderate divergence, the cumulus humilis will be drawn out and sheared off and may be accompanied by dependent stratocumulus.

North of the trade wind maximum, in the Northern Hemisphere, the vertical shear in the wind in the lower layers is usually very great and the trade wind inversion low. The predominant cloud is usually stratocumulus. The clouds will exist in broken patches if the low-level streamline pattern is divergent. Even with weak to moderate convergence at this level, very little cumulus forms; the result is usually an increase in the amount of stratocumulus, west and south of the subtropical anticyclonic centers.

When there is little vertical shear in the lower layers and the low-level streamline map shows weak to moderate convergence south of the trade wind maximum, expect deeper cumulus. The cloud amount will also increase over that expected in divergent flow, but not to a very marked extent. Even with very strong convergence, the cumuliform cloud shows many breaks, the general effect of increased convergence being to increase the height of the convective cloud.

An asymptote of convergence in the low-level streamline field will, if it coincides with a relative minimum in the speed field, be accompanied by a line of large cumulus congestus or cumulonimbus.

Old polar fronts sometimes penetrate into the Tropics. They rarely reach 15°N or S lat in oceanic regions. Such a remnant is often detectable as a line of towering cumulus, and this line between the old air masses also marks a change in cloud form. The old front may, in these cases, appear on the lowest streamline chart as an asymptote of convergence running east to west or northeast to southwest from the neutral point between two middle latitude anticyclones. In most cases, the temperature differences between the old air masses are negligible, even though the cloud line may persist for some time after the air-mass contrasts have disappeared from the soundings.

An asymptote of convergence in the lower streamline field is not necessarily accompanied by a cloud line. If strong speed divergence occurs along part of the line, there may be very little heavy cloud or precipitation accompanying it; furthermore, bad weather may be found at some distance from the line, where speed convergence predominates.

UPPER AIR ANALYSIS

Since the upper winds of the Tropics are far more reliable than upper pressures and temperatures, it is natural to think of upper air analysis at first in terms of streamlines. Such analysis, in fact, has proved useful in day-to-day forecasting; moreover, much of the information we now possess about the tropical atmosphere aloft has been deduced from this method. However, we must consider the pressure and temperature field also. Any numerical values must come from working with the fields of pressure and temperature. Moreover, pressure pattern (D) flying appears to be as feasible over the tropical oceans, at least to latitudes 10° to 15°N and S, as it is in higher latitudes. Thus, for operation of jet aircraft in the high troposphere, a 200-mb analysis is indispensable. Since the fuel supply of jet aircraft is also very sensitive with respect to tempeature—especially during ascent and descent—a close estimate of the temperature distribution is a further requirement.

The upper troughs and ridges of the Tropics usually reach their greatest intensity near 200 millibars, which makes this surface a suitable level for high tropospheric analysis in low latitudes. Also, differential (thickness) analysis is requisite to stabilize the contour analysis of upper isobaric surfaces.

It is impractical to make a thickness analysis for the whole layer from 1,000 to 200 millibars because many disturbances and their associated temperature fields occupy only the lower or the upper troposphere; combination may remove the most significant map features. Therefore it is logical to break the whole layer into two parts and carry out the upper contour analysis in two steps. The 500-mb surface is the most convenient intermediate level, though it is not an ideal surface for tropical upper air analysis because it is situated in the layer intermediate between the disturbances of low and high troposphere. We shall here operate with the layers from 1,000 to 500 millibars and from 500 to 200 millibars. The discussion is limited to latitudes poleward of 10°. Experience still closer to the Equator is too limited to warrant treatment.

General Considerations

Several rules guide the analysis: (1) The upper contours should be drawn parallel to the upper winds when the wind is 10 knots or more. (2) The vertical shear vectors through the two layers we are considering should indicate the orientation of the mean temperature (thickness) field though not necessarily the magnitude of the mean temperature gradient. This applies especially to the layer from 500 to 200 millibars. (3) Since the mean temperature of a layer several hundred millibars thick is nearly conservative from day to day, the thickness fields of two succeeding days should be readily comparable. Regions of relatively warm and cold air should be traceable from day to day without difficulty and the intensity of centers should not change drastically without good reason. In particular, one should not find huge areas where colder or warmer air, not present previously, appears suddenly. When this happens, analysis errors should always be suspected. (4) Heights of upper isobaric surfaces and thicknesses between surfaces vary in a limited range in each area and season. Normals and extremes can be established by plotting frequency distributions. This will help the analyst to avoid values outside the observed ranges and to realize when he has drawn a higher or low value with the permissible range. We have seen earlier that such high or low values are associated with definite types of flow patterns.

Analysis Procedure

The usual approach to analysis when the thickness method is used, is to first draw the upper contour charts, then compute the height differences between the isobaric surfaces, and inspect the result as to pattern and change from the preceding day. Assuming the 500-mb analysis is finished, draw next the 1,000-mb chart. Then obtain the thickness either graphically or plot a chart containing thickness values and wind shears at the observation stations along with thicknesses at all 5° lat/long intersections over the oceans. Although the latter procedure is preferable, it is more likely that we will use the graphic approach, primarily because of limitations on our time. Depending on the area and season, the interval for analysis is 50 or 100 feet (15 or 30 meters). We then inspect our charts carefully and make the necessary adjustments.

This process is then repeated for the 500- to 200-mb layer. In addition to larger and more definitive wind shifts, the Aerographer's Mate can use here the fact that a high correlation exists between the height of the 200-mb surfaces and the thickness of the layer from 500 to 200 millibars.

As analyst, the Aerographer's Mate can correlate the cloud and weather chart described earlier with the upper flow patterns and thickness changes. In a zone where the cloud reports indicate definite subsidence, cooling from the previous day cannot be observed unless a cold center is being advected into the area; the latter then should be losing intensity with time. A deepening upper cyclone must either have cloudiness on the inflow side into the cyclone—which is often incompatible with the dynamics of the situation—or colder air must be brought in from outside the system. Otherwise there is no physical process except radiation which could achieve the central cooling which must be present to account for 200-mb height decreases.

The relation between anticyclonic flow and weather is just as important. High tropospheric anticyclones can increase in intensity, coupled with warming in middle and lower troposphere, both in areas of ascent and descent. In the latter case we are dealing with the well-known dynamic high. If a 200-mb high strengthens, however, when much cloudiness is present, either aloft or throughout the troposphere, this can only mean that latent heat of condensation is being liberated in such a way as to produce warming aloft. The heat liberated can be partly converted to kinetic energy in such cases and there is then the possible beginning of a storm.

Estimating Wind Flow From Satellite Pictures

Clouds over the Tropics, viewed from satellite altitude, reveal many features of the flow. The distribution of widespread cloud systems has definite relationships to major trough and ridge positions. Wind estimates for both upper and lower tropospheric level can be obtained from an analysis of cumulus and cirrus cloud formations. These can be obtained by using interpretation of singular cloud pictures or on actual measurement of cloud motion from a series of pictures.

Two of the most useful cloud patterns used for determining upper level flow are plumes from the tops of cumulonimbus clouds, and cirrus cloud shields associated with subtropical jet streams.

Vertical wind shear through an atmospheric layer containing clouds is one of the prime factors governing the shape of cloud formation. Most of the cloud lines and bands seen in satellite pictures are oriented parallel to the thermal wind through the layer from cloud base to cloud top. Under some circumstances, the direction of shear is the same, or nearly the same, as the direction of motion of the air containing the cloud. This is the case if: (1) the wind changes speed with height but not direction; (2) the wind changes direction with height but the winds at the top of a cumulonimbus are much stronger than those near its base.

When these conditions occur, it is possible to estimate the wind direction directly from the orientation of the cumulonimbus plumes. These conditions are common in tropical regions where the speed of the wind at the level of cumulonimbus anvils is several times greater than the mean wind speed of the wind through the lower portion of the cumulonimbus cloud. In these cases wind estimates based on orientation and dimensions of cirrus blow-off are quite accurate and closely approximate the winds near the cloud top.

There are two conditions which can cause considerable error in wind direction estimates derived from cumulonimbus plumes: (1) light winds at the cirrus level; (2) a large difference between the direction of the shear through the convective layer and the direction of the wind at the top of the layer.

When the speed is light (less than 20 knots), the wind direction is variable causing the correlation between plume orientation and wind direction to be less reliable. If there is a large difference between the direction of the shear through the convective layer and the direction at the top of the layer, the plumes will be parallel to the shear but not to the wind at the layer top. This condition is common in frontal areas where temperature advection is strong, but occurs less frequently in the Tropics.

UPPER LEVEL FLOW FROM CIRRUS CLOUD SHIELD.—It is common to observe huge masses of cirriform clouds extending poleward from the tropics. These clouds occur in advance of 200-mb troughs and represent a poleward transport of momentum and high level moisture from the tropical region. A wind maximum is usually located on the poleward edge of these cloud formations. As with strong jet streams, the direction of the upper level wind parallels the cloud edge. Because the air is moving away from the equator and accelerating, it is crossing the contour pattern toward lower pressure. For this reason, the general orientation of the cloud shield and the striations within it can differ as much as 15 degrees from that of the upper tropospheric contours. When this is taken into account, cirrus formations of this type provide good estimates of wind direction but less precise information as to wind speed. Figure 12-33 shows a typical cirrus cloud shield. The double-shafted, white arrows (B to C) represent the 200-mb wind maximum suggested by the cirrus cloud edge. The black, single-shafted arrows represent the 200-mb flow based on individual cirrus elements that lie within the larger scale formation. The vortical cloud pattern (D) is associated with a cut-off low. Southeast of D, a cirriform cloud shield extends northeast from A in advance of an upper level trough associated with the vortex. There are numerous small-scale lines of clouds (E to F) which are oriented approximately perpendicular to the wind direction. These lines are believed to be caused by horizontal shear and should not be confused with cumulonimbus plumes which are oriented parallel to the vertical shear.

LOW LEVEL WINDS FROM CUMULUS CLOUD LINES.—Research has shown that many cumulus cloud lines observed between 10N and 10S over the Atlantic and eastern Pacific Oceans are oriented approximately parallel to the surface wind direction. Because the orientation of cloud lines relates to the shear, these lines may depart considerably from the surface wind direction. Therefore, a great deal of caution must be exercised when using them to define low-level wind flow. There are no hard, fast rules which permit positive identification of the particular cloud lines which are approximately parallel to the surface wind. It has been observed that cloud lines which are very long, very narrow, and either wavy, zig-zag, or have knots (wide places along a line), are the type most often parallel to the surface wind. In figure 12-34, cloud lines are shown which are approximately parallel to the wind flow. Surface wind reports are entered on the picture as white arrows.

Upper-Level Streamline Analysis

All the analysis rules applicable to the low levels also apply to the higher parts of the troposphere. In the upper troposphere, in the vicinity of the 200-mb level or the 40,000-ft (12,190 meters) level, you find more often than not that your data has diminished to less than half that available for the low-level streamlines, and in view of this you find that the qualitative streamline analysis method is more suitable for this level in most cases. The streamline analysis at this level is used primarily for the detection of cyclonic and anticyclonic flows, and, of course, you look most for the signs of anticyclonic outflow at the 200-mb level, for it is this outflow that is the first clue to tropical cyclone development in the upper air.

Before you finalize your analysis, you should look over your chart objectively and ask yourself the following questions:

1. Do the streamlines conform to the general flow pattern?

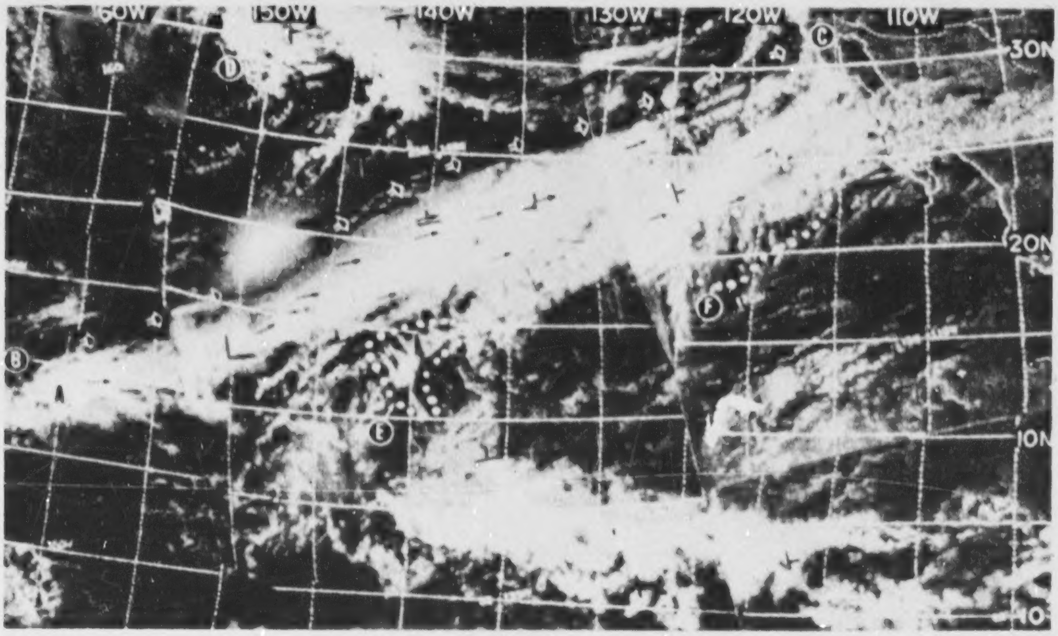

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Figure 12-33.—A typical cirrus cloud shield.

2. Have I justified throwing out every wind that I could not draw to?
3. If I cannot justify discarding the wind, is there any way I can draw to it?
4. Have I drawn the streamlines parallel to the plotted winds rather than at an angle to winds?
5. Is the chart consistent with other levels?
6. Is the chart consistent with history?
7. Are the streamlines and isotachs consistent?
8. Have I drawn any unnecessary lines?
9. Have I given more weight than necessary to light and variable winds?

TROPICAL FORECASTING

Forecasting in the Tropics is a difficult problem, necessitating a good meteorological and physical background, vast amounts of climatological knowledge, a keen mind's eye which can differentiate the minutest deviation in a mass of nearly homogeneous data, and last, but not least, diligence and dedication in the approach to the forecast.

The types of forecasts in the Tropics are the same as anywhere, in that you encounter flight forecasts, route and terminal; operational forecasts, and general or fleet forecasts (RATT broadcast form); weather clearances; local area forecasts and advices; and destructive weather forecasts and warnings.

Your concern in this chapter is with the local area forecasts, forecasting the ITCZ, tropical waves, and destructive weather forecasts and warnings for the reasons that the other forecasts will basically depend on the same general techniques and approach for their formulation as the local area forecast. The treatment of destructive weather warnings and forecasts is limited to tropical storms and cyclones, because tornadoes are largely non-existent in the Tropics and

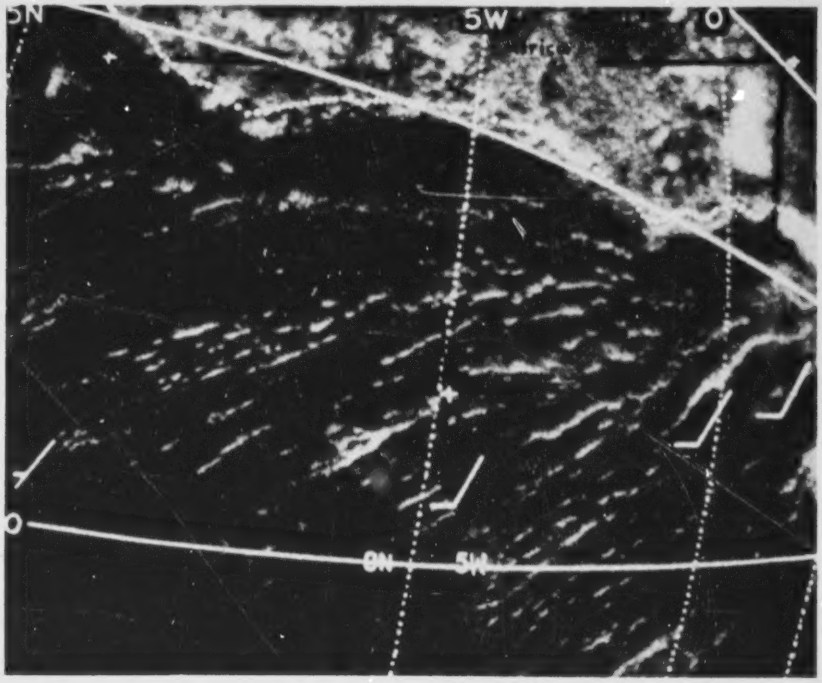

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Figure 12-34.—Low level clouds off the coast of Africa.

thunderstorms are covered in chapters 11 and 12 of this course.

LOCAL AREA FORECASTS

The importance of the local area and the general area climatology has been pointed out at various points in this chapter. It is in the preparation of the local area forecast that this knowledge will be most beneficial.

During the analysis of the various charts, most forecasters form a mental image of the forecast charts and perceive certain fundamental ideas of the local weather (and the weather anywhere in the area of responsibility, for that matter) for the next 24 or 48 hours. The climatology of the analysis and forecast area serves as a guide in the analysis and in the mental prognostication. The next step in the procedure is to expand and refine the mental image formed so far.

The ideal approach to a local area forecast is to prognosticate the upper air analysis first, usually the 500-mb chart, from which (with other things in consideration, as for instance, streamline analysis, weather distribution charts, and time sections) the surface chart is prognosticated. This prognostic surface chart is then used as a basis for the local area forecast, again with these other factors considered and correlated, and also correlated to the climatology of the local area.

Many times this routine for various reasons cannot be completed in its entirety, and the forecast must be prepared anyway. In that case the approach is still the same, in that ALL available data are considered and the best is done with what data are available.

TROPICAL CYCLONE FORECASTING

There is at present no one formal procedure for forecasting the development and movement of tropical cyclones. This can be understood when one considers the enormous complexity of the problem, the sparsity of data in the oceanic tropical regions compared to that available in the well-developed and highly populated continents, and the vanishing of ship reports from the area of a tropical cyclone as soon as the first warning is broadcast. There is also the regional influence to consider, and this influence cannot be slighted. A very obvious consequence of regional influence can be demonstrated when you compare the North Atlantic area with the North Pacific area. The North Pacific has almost twice the tropical water area and also better than double the average number of tropical cyclones per annum.

Forecasting tropical cyclones evolves into the following problems: formation, detection, location, intensification, movement, recurvature, and decay.

The factors that enter into the forecast preparation are mainly dynamical (relating to the energy or physical forces in motion), but there are also important thermodynamic influences. For instance, tropical storms will not develop in air that is a little drier, and generally also slightly cooler, in the surface layers than the air normally present over the western portions of the tropical oceans in summer. This holds true if air with a trade inversion and upper dry layer are present. Surface ship temperature and dewpoint reports, along with upper air data, are extremely valuable in determining whether the surface layers are truly tropical or whether they contain old polar air not yet completely transformed. Careful observations from ground and aircraft observers on convective activity indicate whether or not there is a lid against upward penetration of the cumuli.

Dynamically storms (existing and potential) are subject to influences from the surrounding areas. In the Tropics, we usually encounter two layers in the troposphere, and these levels have very different characteristics. Mostly, a steady trade blows in the lower levels (surface to 500 mb), while a succession of large cyclonic and anticyclonic vortices are present in the upper troposphere in the 400- to 150-mb strata. In some regions, however, (for instance between the Mariana Islands and the South China Sea in midsummer) intense eddy activity takes place also near the ground. It is not possible to deduce from charts drawn in the lower layers what is taking place above 500 millibars. Therefore, it is necessary to keep track of events in both layers.

Aside from the surface chart, the 700-mb level chart is very helpful in determining low-level flow patterns. The 200-mb level is representative of the upper layer, whereas the 500-mb level, often located in the transition zone, is of much less use in tropical than in extratropical forecasting.

However, forecast considerations should not be limited to the two layers in the Tropics. Middle latitude weather and changes also influence and share in the control of weather changes in low latitudes. The position and movement of troughs and ridges in the westerlies affect both formation and motion of tropical storms. Therefore, middle latitude analysis is a requisite. Since forecasts generally run from 1 to 3 days, this should be a hemispheric analysis.

The climatological approach to the forecast problem should also be taken into consideration. It is obvious that a forecaster must be familiar with his region and the seasonal changes: for instance, areas in which storms tend to form or not form for any given month; mean storm tracks and the scattering of individual tracks about the mean. For the Atlantic area an elaborate set of mean tracks, percentage of frequency of motion and median speeds for 5° latitude squares, and regions of development and their subsequent effect on east coast weather are available in the U.S. Department of Commerce, Forecasting Guide No. 3, Hurricane Forecasting. Tracks and data on Pacific storms are also available in other publications. These maps and charts reveal that the "climatological" approach gives some information, but it cannot be relied on in any area, or in the case of any particular storm, to the exclusion of synoptic indications for the forecast. These probability considerations are useful mainly for long-range planning and when data are missing. Therefore, we reach the conclusion that climatological information should be treated as a weighting factor to be included after synoptic data.

Formation

A full blown typhoon/hurricane cannot be forecast as such when there are no indications of any type of irregularity on the charts and aids used in tropical analysis. Only when the incipient stage first appears among the data can we begin to think in terms of an actual typhoon/hurricane, and often not even then. For the most part the forecast from "tropical low" to typhoon/hurricane" is a matter of step-by-step progression. Assume the term "formation" means formation of a "potential" typhoon/hurricane.

Empirical rules or checks have evolved through the years which enjoy a measure of reliability to warrant their use (of course, the more signs point to formation, the more likely formation will be).

The synoptic conditions favorable for development were given in a previous section of this chapter. In the following list, no attempt is made to separate the surface from the upper air indications as the two are most often occurring simultaneously and are interrelated.

1. Marked cyclonic turning in the wind field.
2. Shift in the low-level wind directions when easterlies are normally present. Wind-speed must be 10 knots or more.
3. Greater than normal cloudiness, rainfall, and pressure falls. Cloudiness should increase in vertical as well as horizontal extent.
4. Sea level pressures lower than normal. The value is dependent upon the region analyzed; however, a value greater than 3 millibars is the normal criterion.
5. Easterly wind speeds 25 percent and more above normal in a limited area, especially when the flow is cyclonic.
6. Temperatures above normal at sea level, generally 26°C (79°F) or more in the lower layers.
7. Moisture above normal at all levels.
8. Westerlies greater than average north of the latitude of the seasonal maximum.
9. Easterlies weaker than average in a wide zone.
10. Easterlies decreasing with height.
11. Easterlies over or approaching 40,000 feet.
12. Latitude of the subtropical ridge higher than normal above the 500-mb level.
13. There is evidence of a fracture of a trough aloft (at 200 millibars).
14. Long waves slowly progressive.
15. A zone of heavy convection is present, indicating the absence of the trade inversion.
16. Surface pressure gradient north and
17. Disturbance has relative motion toward upper ridge at or above 400 millibars.

Other indications of development which are useful are sea swell and tide observations. Swell will have a period less than average and an amplitude greater than average. The normal swell frequency is 8 per minute in the Atlantic and 14 per minute in the Gulf of Mexico. Hurricane winds set up swells with a period that can decrease to four per minute. Swells will approach the observer approximately from the direction in which the storm is located. The swell height is an indication of the storm's intensity, especially when the swells have not encountered shallow water before reaching shore.

Abnormally high tides along broad coastlines and along shores of partially enclosed water bodies with the highest tides usually to the right of the path looking downstream, are also a good indication.

Detection

We already discussed how the meteorological satellites have greatly aided in the detection of tropical disturbances, especially in the early life cycle. It is important for meteorologists, analysts, and forecasters to be able to effectively interpret these pictures to achieve the maximum information from them. Through proper interpretation of the pictures determination of size, movement, extent of coverage, and approximate surface winds can be made. Satellites have become the most important method of detection of disturbances.

Aircraft reconnaissance plays an important role in detection also. This has become even more important when used in conjunction with the data from satellite pictures. Areas of suspicious cloud structure can be investigated by aircraft whose crews include trained meteorologists and observers. These reconnaissance flights can provide on-station data by orbiting or penetration, either high- or low-level, that would not be available otherwise.

Another method of detecting tropical disturbances is through the use of radar. Present radar ranges extend about 300 miles and can scan an area approximately 300,000 square miles.

The Navy has also developed several different types of floating weather stations to report various meteorological elements. These stations send out measured data automatically or upon query.

Location

The problems of formation, detection, and location are in reality a single three-in-one problem. One is dependent on the other.

In the case of satellite pictures, reconnaissance, and radar detection, the location is fairly certain, barring navigational errors. If detection is made through the analysis, the exact location is more difficult to ascertain in the incipient stage of the storm, especially when the analysis is diffuse. The exact location should be decided upon only after the most intensive study of the data. The analyst should be prepared to revise his decision in the face of developments which are more conclusive.

Seismic aids have also been used to some extent in detection of tropical cyclones; however, very little is published on the subject. Discussion of this aid cannot go beyond mention that it has been used.

Intensitifation

Only when easterlies extend vertically to 25,000 feet or more at the latitude of the vortex is intensification possible. This most frequently occurs when the subtropical ridge lies poleward of its normal position for the season. Additional indicators are as follows:

1. Movement is less than 13 knots.
2. The disturbance is decelerating or moving at constant speed.
3. The system has a northward component of motion.
4. A migratory anticyclone passes to the north of the storm center.
5. Intensification occurs when the cyclone passes under an upper-level trough or cyclone provided that there is relative motion between the two. There is some indication that intensification does not take place when the two remain superimposed.
6. Poleward movment of the cyclone is favorable for intensification; equatorward motion is not.
7. Intensification occurs only in areas where the sea surface temperature is 79°F or greater, with a high moisture content at all levels. (This rule has not been thoroughly tested.)
8. When long waves are slowly progressive. (Applicable to genesis also.)
9. The trade inversion must be absent; convection deep.
10. Other factors being equal, deepening will occur more rapidly in higher than in lower latitudes. (Coriolis force is stronger.)
11. When a storm moves over land, the intensity will immediately diminish. The expected amount of decrease in wind speeds can be .30 to 50 percent for storms with winds of 65 knots or more; and 15 to 30 percent for storms with less than 65 knots, if the terrain is flat; there is more decrease in each case if the terrain is rough.

Once an intense tropical cyclone has formed, there will be further changes in its intensity in the course of its motion within the Tropics and during recurvature. The forecaster should consider these changes in connection with the predicted path of movement.

Movement

Tropical cyclones usually move with a direction and a speed which closely approximate the tropospheric current which surrounds them.

Logically, therefore, charts of the mean flow of the troposphere should be used as a basis for predicting the movement of tropical cyclones, but lack of observations generally precludes this approach.

In the mean, there is a tendency for tropical cyclones to follow a hyperbolic or parabolic curve away from the Equator; however, departures from this type of track are frequent and of great variety.

Tropical cyclones move toward the greatest surface pressure falls and toward the area where the surface pressure falls increase fastest with time. Calculation is necessary for this rule to be used.

Numerous theories have been advanced to explain the cyclone tracks of the past and to predict those of the future. Observational data have never been sufficient to prove or disprove most of them. A few theoretical concepts have found limited application, but for the most part the forecaster must rely upon empirical knowledge and extrapolation when predicting the movement of tropical cyclones. In this section are discussed some of the current theories and rules on forecasting these tracks; and one objective method found successful in a large number of cases is taken up in detail.

Tropical storms move under the influence and are a result of both external and internal forces. The external force is due to the air current that surrounds the storms on all sides and carries them along. The internal forces appear to produce a tendency for a northward displacement of the storm which probably is proportional to the intensity of the storm, a westward displacement that decreases as latitude increases; and a periodic oscillation about a mean track.

INITIAL MOVEMENT OF INTENSE CYCLONES.–The movement of cyclones that are undergoing or have just completed intensification initially (about 24 hours) will be as follows:

1. Storms developing in westward moving wave troughs in the easterlies move toward the west and also with a poleward component given by an angle of approximately 20° to the right of the axis of the trough looking down-stream. (See fig. 12-35.)

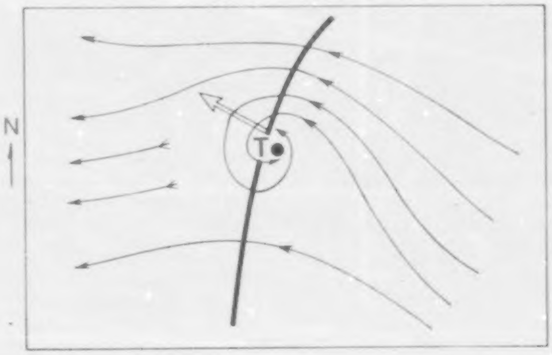

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Figure 12-35.–Illustration of initial movement of tropical cyclones forming in a wave trough in the easterlies.

2. The motion of storms developing from preexisting vortices can be extrapolated from the track of these vortices.

3. Storms developing on the eastern edge of polar troughs initially move up the trough (NW to NE.

4. Pacific storms forming on equatorial shearlines usually are most active in the southern and southwestern portions at the start and they move slowly northeast. After 1 to 2 days, the influence of the trades becomes dominant and the storms turn back toward a direction ranging from west to north. (This appears to be true mainly for cyclones deepening near or west of 130°E.)

5. If a storm is discovered without information as to its past movement and the upper air current, it is best to start it along the climatological mean track and then secure the data necessary to determine the steering current.

EXTRAPOLATION.–At present, the most practical prognostic technique used by the tropical meteorologist consists of extrapolating the past movement of the synoptic features on his chart into the future. The past track of a cyclone represents the integrated effects of the steering forces acting upon it. Accelerations and changes in course are the results of the changes in these steering forces. The effect of these forces can be examined and extrapolated directly from the past position of the cyclone.

Before applying the extrapolation technique, the forecaster must attempt to smooth out the minor irregularities in the past track. The first step in the use of extrapolation is to determine the mean direction and speed of the cyclonic center between each two known positions. The next step is to determine the rate of change in direction and speed between successive pairs of fixes. The forecast position thus determined from extrapolation of movement should be smoothed out for minor irregularities and compared to the applicable climatological tracks of cyclones in the area. If large differences exist, the forecast should be completely reexamined. The climatological tracks should receive less weight for short-term forecasts of 6 to 12 hours and more weight for forecasts in excess of 24 hours.

For short-term forecasts (6 to 12 hours) extrapolation is as reliable as any known method for movement.

STEERING.–The movement of a tropical cyclone is determined to a large extent by the direction and speed of the basic current in which it is embedded. This concept appears to work well as long as the cyclone remains small and remains in a deep broad current. By the time a tropical cyclone has reached hurricane intensity, these conditions seldom exist. It then becomes necessary to integrate the winds at all levels through which the cyclone extends and in all quadrants of the storm to determine the effective steering current. Although the principle is not fully understood and has not yet received universal acceptance as a valid rule, it does have practical applications. For example, changes in winds at one or more levels in the area surrounding the cyclone can sometimes be anticipated, and in such cases, a qualitative estimate of the resulting change in the movement of the cyclone can be made. Conversely, when it appears likely that none of the winds in the vicinity of the cyclone will change appreciably during the forecast period, no change in the direction and speed of movement should be anticipated.

Further, this rule applies to situations of nonrecurvature and some cases of recurvature. It is difficult to determine the steering current, since the observed winds represent the combined effects of the basic current and disturbances. As most storms extend into the high troposphere, it is better to calculate a "steering layer" than a steering level, since presumably the wind throughout most of the troposphere influences the storm movement. One writer recommends an integration of the mean flow between the surface and 300 millibars, over a band 8° in latitude centered over the storm. Another writer indicates that for moderate and intense storms the best hurricane steering winds would be found in the layer between 500 and 200 millibars and averaged over a ring extending from 2° to 6° latitude from the storm center.

Other practical applications of the steering concept to short-range tropical cyclone motion use differing approaches in attempting to measure the basic current. One is by streamline analysis of successive levels to find a height at which the vortical circulation diminishes to a point such that the winds are supposedly representative of the undisturbed flow. Another method is to take vector averages of reconnaissance winds near the zone of strongest winds in the storm.

1. Use of Observed Winds Aloft for Steering. When sufficient data are available, it has been found that the use of streamline analysis of successive levels usually gives valuable indications of tropical storm movement for as much as 24 hours in advance. However, since wind observations are usually scarce in the vicinity of a hurricane, their analysis is necessarily rather subjective. Some forecasters claim dependable results in using this concept when data were available to high levels near the storm. The technique is not based on the assumption that wind at any single level is responsible for steering the storm, since the forces controlling movement are active through a deep layer of the atmosphere. However, as successive levels are analyzed, a level is found at which the closed cyclonic circulation of the storm virtually disappears. This STEERING level coincides with the top of the warm vortex and varies in height with different stages and intensities of the storm. It may be located as low as 20,000 feet or in the case of a large mature storm as high as 50,000 feet.

It has been found in analysis that most weight should be given to the winds in advance of the storm within a radius of 200 to 300 miles in preference to those in the rear quadrants. The hurricane generally moves with a speed of 60 to 80 percent of the current at the steering level.

The direction of movement is not always exactly parallel to the steering current, but has a component toward high pressure which varies inversely with the speed of the current, ranging from al~~most~~ 0° with rapid movement to as much as 20° with speeds under 20 knots. In westward moving storms, a component of motion toward high pressure could result from the poleward acceleration arising from the variation of the Coriolis parameter across the width of the storm. This would indicate that, to the extent that this effect accounts for the component of motion toward high pressure, northward moving storms would fit the direction of the steering current more closely than westward moving ones. The tendency for poleward drift would be added to the speed of forward motion in case of a northward moving storm so that it would approach more closely the speed of the steering current. Empirical evidence supports this hypothesis.

Corrections for both direction and rate of movement should be made when this is indicated by the windflow downstream in the region into which the storm will be moving. For prediction beyond several hours, changes in the flow pattern for a considerable distance from the storm must be anticipated. It should also be remembered that intensification or decay of a storm may call for use of a higher or lower level, respectively, to estimate the future steering current.

2. Thermal Steering. A number of efforts have been made to correlate hurricane movement with thermal patterns. For example, Simpson suggests that a storm will move along tongues of warm air in the layer from 700 to 500 millibars that often extend 1,000 miles ahead of the storm. The orientation of the axis of the tongue then may be regarded as a reliable indicator of storm movement for the next 24 hours. A considerable difficulty in applying this technique is created by the fact that the warm tongue sometimes has more than one branch and it is questionable as to which is the major axis. An example of Simpson's method of warm tongue steering method can be found in the U.S. Department of Commerce, Forecasting Guide No. 3, Hurricane Forecasting.

Recurvature

One of the fundamental problems of forecasting the movement of tropical cyclones is that of recurvature. Will the cyclone move along a relatively straight line until it dissipates, or will it follow a track which curves poleward and eastward? When recurvature is expected, the forecaster must next decide where and when it will take place. Then, he is faced with the problem of forecasting the radius of the curved track. Even after the cyclone has begun to recurve, there are a great variety of paths that it may take. At any point, it may change course sharply.

The most common recurvature situation arises when an extratropical trough approaches a storm from the west or when the storm moves west to northwest toward a stationary or slowly moving trough. Some of the indicators on possible recurvature are as follows:

1. If the base of the polar westerlies lowers (to 15,000 to 20,000 feet) west of the storm at its latitude and remains in this position, recurvature may then be expected to occur.
2. However, if there is the building of a dynamic high or an eastward movement of this high to the rear of the advancing trough, and the westerlies dissipate in the low latitudes, the storm will move past the trough to its south and continue its westward path.
3. Rule 2 also holds true in cases where the polar trough moves from the west against a blocking high. The higher latitude portion of the trough continues to move eastward while the southern segment of the trough is retarded and is no longer connected with the upper portion of the trough.
4. Recurvature may be expected when an anchor trough is about 500 miles west of the storm and when the forward edge of the westerlies is from 500 to 700 miles to the west. Correlated with this parameter R. J. Shafer found that a spot value of the thickness between 850 to 500 millibars 7.5 degrees of latitude to

the northwest of the storm was one of the most valuable of all parameters on the recurvature or as an indication of future movement of Atlantic hurricanes. This thickness reflects the relative strength of cold troughs to the west essential for recurvature. Low values of thickness 14,000 feet (approximately 4,270 meters) or less, almost always indicate recurvature and high values, 14,200 feet (4,330 meters) or more, generally indicate continued westward motion.

5. The major trough west of the storm (in the westerlies) is slowly progressive.

6. Long waves are stationary or slowly progressive.

7. There is a rapid succession of minor troughs aloft.

8. Climatological mean track indicates recurvature (use with caution).

9. When the neutral point at the southern extremity of the trough in the westerlies at the 500-mb level lies at or equatorward of the latitude of the cyclone, recurvature into the rough will usually occur. In this situation, the cyclone would normally be under the influence of southerly winds from the upper limits to a level well below the 500-mb level while approaching trough.

10. When the subtropical ridge at the 500-mb level is broad and consists of large anticyclones, recurvature usually occurs. This case represents a low index situation in which the cyclone remains under the influence of a single, large, slow-moving anticyclone for a relatively long time.

11. Weak troughs between two separate subtropical high cells. Sometimes the tropical storms move northward through very weak breaks in subtropical highs.

NONRECURVATURE.–The following flow patterns are associated with nonrecurvature:

1. Strong subtropical anticyclone or ridge to the north of the storm with the mean trough in the westerlies located far to the west of the longitude of the storm. If this pattern develops strongly over the western oceans or continents, a storm will generally be driven inland and dissipate before it recurves into the western end of the ridge.

2. Flat (small-amplitude waves) westerlies at latitudes near or north of the normal position. A narrow subtropical ridge separates the westerlies from the tropical trough. In many cases the mean trough in the westerlies may be located at the same longitude of the storm.

MOTION DURING RECURVATURE.–The following rules apply during recurvature only:

1. When the radius of recurvature of a storm is greater than 300 miles, it will not decelerate and may even accelerate. The storm will slow down if the radius of recurvature is less than 300 miles. In general when the radius of recurvature is large, it is usually very uniform. A small radius will occur along a brief portion of the track. (See fig. 12-36.)

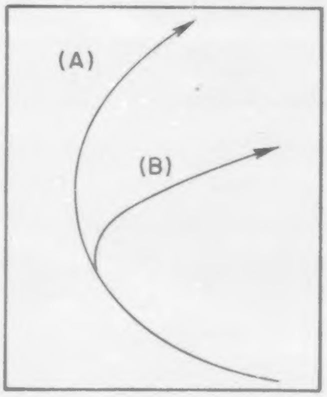

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Figure 12-36.–(A) Recurvature at a constant speed; (B) first decelerating and then accelerating.

2. A large radius of recurvature is to be expected if the high northeast of the tropical cyclone has a vertical axis (fig. 12-37 (A)). This will occur when long waves are stationary.

3. When the high slopes south to southeast with height (fig. 12-37 (B)), the cyclone is transferred rapidly from the influence of upper easterlies to that of the upper westerlies and the track has a short bend. This occurs when long waves are progressive.

The following rules refer to short term (24 hours or less) forecasting:

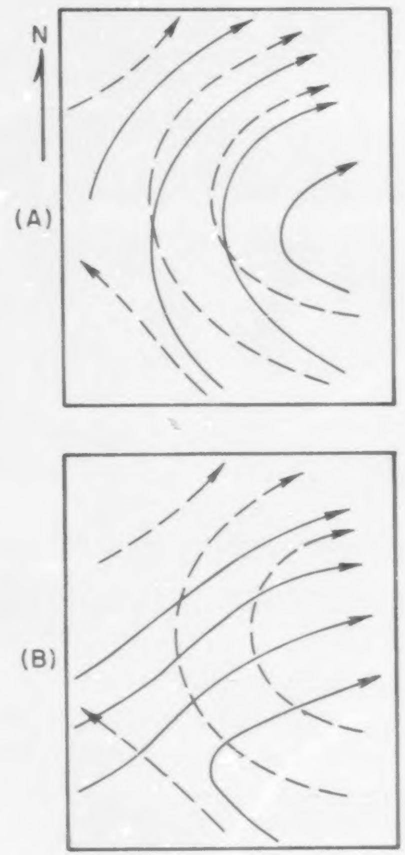

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Figure 12-37.—Surface flow pattern (dashed line) and 300- to 200-mb flow pattern (solid lines). (A) Corresponds to track (A) in figure 12-37; (B) corresponds to track (B) in figure 12-37.

1. Tropical cyclones move toward the area of greatest surface fall (12- to 24-hour pressure change).
2. Tropical cyclones move toward the area where the surface pressure falls increase most rapidly with time. For calculation 3-hourly reports are necessary. Take, for example, the 24-hour pressure change at 1800 and subtract it from the 24-hour pressure change at 1500. This gives the acceleration of pressure fall. If this quantity is computed for a network of stations and isolines are drawn for a suitable interval, the negative center helps to pinpoint the expected cyclone position. This rule is especially helpful when a storm is just offshore to determine the precise place of entry for intervals of 12 hours or less.

Changes in Intensity During Movement Out of the Tropics

During movement out of the Tropics the storm comes under influences different from those in its birthplace and tropical path. The following rules and observations can be helpful in forecasting the storm's changes in intensity.

1. A storm drifting slowly northward with slight east or west component will preserve its tropical characteristics.
2. Storms moving northward into a frontal area or area of strong temperature gradients usually become extratropical storms. In this case the concentrated center dies out rapidly while the area of gale winds and precipitation expands. The closed circulation aloft gives way to a wave pattern and the storm accelerates to the northeast.
3. If the tropical cyclone recurves into a trough containing a deep slowly moving surface low, it normally will overtake this low and combine with it. Temporary intensification may result.
4. If the tropical cyclone recurves into a trough containing an upper cold core low, it appears to move into the upper low in most cases.

Short-Range Prediction by Objective Techniques

As in all so-called objective techniques no one method will serve to produce an accurate forecast for all tropical storms or extratropical systems. In the following section several techniques which have been developed in recent years are discussed. It should be mentioned that the objective technique should not be considered as the ultimate forecast and that all other rules, empirical relationships, and synoptic indications should be integrated into the final forecast. Wholly inaccurate forecasts may result if the objective method were the only one considered in making the forecast.

STATISTICAL METHODS. One of the most prominent and widely used of these methods was devised by Veigas-Miller to predict the 24-hour displacement of hurricanes based primarily on the latest sea level pressure distribution. Sea level pressures were used rather than upper air data due primarily to the longer available record of sea level data and also because of the advantage of denser areas and time coverage and more rapid availability of the data after observation time. In addition to the sea level pressure field this method also incorporates the past 24-hour motion of the storm and climatological aspects of the storm.

Pressure values are read from predetermined points located at intersections of latitude and longitude lines with values divisible by 5. Two different sets of equations are used: one set for a northerly zone (between latitudes 27.6° and 40.0°N and longitudes from 65.0° to 100.0°W). The southerly zone encompasses the same longitudes as the northerly zone, but the altitudes are for 17.5° and 27.5°N.

This method was tested by Veigas and Miller on 125 independent cases, about equally distributed between the two zones, during the years 1924-27 and 1954-56. The average vector errors in 24-hour forecast position were about 150 nautical miles for the northerly zone and 95 nautical miles for the southerly zone.

Recently the University of Hawaii has modified this set of equations so that they will be applicable in the Pacific area. A complete test has not been made and at the time of this writing, the equations have not been distributed for operational use.

Full details, with an actual example of the procedure for use of this method; may be found in the U.S. Department of Commerce, Forecasting Guide No. 3, Hurricane Forecasting.

30-HOUR MOVEMENT OF CERTAIN ATLANTIC HURRICANES. R.J. Shafer has developed an objective method for determining the 30-hour movement of hurricanes by use of sea level data over the ocean areas and upper air data over land area. Motion is described in two components: zonal and meridional. Westerly motion is determined by consideration of the component of the sea level pressure gradient surrounding the hurricane and is opposed graphically by the mean temperature field between 850 and 500 millibars. The correlation of these parameters is modified by extrapolation and the geographic locations. Meridional motion is similarly predicted by sea level and thickness parameters modified by extrapolation. The meridional and zonal computations are then combined into the final 30-hour forecast.

In a test of dependent and independent data it was found that some 85 percent of the storms predicted in the 31 sample cases fell within 2° of the predicted position.

The complete details of this method can be found in GRD Scientific Report No. 1, Contract No. AF19(604)-2073, Further Studies in the Development of Short-Range Weather Prediction Techniques, April 1958.

MOVEMENT OF TYPHOONS. Griffith Wang of the Civil Air Transport Service, Taiwan, developed a method for objectively predicting the movement of typhoons in the western Pacific. The method is titled "A Method in Regression Equations for Forecasting the Movement of Typhoons." The equation utilizes 700-mb data and is based on the following criteria:

1. The 700-mb contour height and its tendency 10° lat north of the typhoon center.
2. The 700-mb contour height and tendency 10° lat from the typhoon center and 90° to the right of its path of motion.
3. The 700-mb contour height and its tendency 10° lat from the typhoon center and 90° to the left of its path of motion.
4. The intensity and the orientation of the major axis of the subtropical anticyclone which steers the movement of the typhoon.

Percentage of frequency of direction of movement and speed tables are provided for a rough first approximation of the movement of the typhoon.

This method, as well as all other methods based on a single chart, is dependent upon a good network of reports and a good analysis. A full test of the value of this method has not been made, but in limited dependent and independent data cases tested it appears to have a good verification and provides another useful tool in the integrated forecast.

Full details on the procedure and application of this method can be found in the Bulletin of the American Meteorological Society, Vol. 41, No. 3, March 1960.

USE OF THE GEOSTROPHIC WIND FOR STEERING. The expansion of the aircraft reconnaissance reports have made it practical to carry out more detailed analyses of constant pressure surface over the tropical storm belt and make use of a forecast based on geostrophic components at that level. This technique, developed by Riehl, Haggard, and Sanborn, and issued as an NA publication (Objective Prediction of 24-Hour Tropical Cyclone Movement) utilizes this steering concept. The technique makes use of 500-mb height averages alongside of a rectangular grid approximately centered on the storm. The grid is 15° long, centered at the initial longitude of the storm and between 10° and 15° lat with the southern end fixed at a distance of 5° lat south of the latitude of the storm center. The more northward extension of the grid is used for storms found to be moving more rapidly northward. The relatively small size of the grid indicates that tropical cyclone motion for 24 hours is determined to a great extent by circulation features closely bordering the storm, and that only the average features outside this area will not greatly affect its movement within the time interval. This method is discussed in detail later in this chapter. The 500-mb chart is the basic chart for computations.

Another method which uses the steering concept is A Comparison of Hurricane Steering Levels by B. I. Miller and P. L. Moore of the United States Weather Bureau Hurricane Forecasting Service. In their study it was found that the standard rule for steering tropical cyclones (the movement of the storm from about 10° to 20° to the right of the current flowing over the top of the core) was only reliable prior to recurvature and that storms after recurvature frequently move to the left of the steering current. Operating on the premise that the motion of the tropical storm is not governed by forces acting at any one level, their study encompasses three levels, the 700-, 500-, and 300-mb levels. They found that the 700- and 500-mb charts were about equal in forecasting hurricane motion. In the final analysis, the 700-mb level was selected and combined with the previous 12-hour motion of the storm. From their study a slightly better verification of predicted tracks of hurricanes resulted than from use of sea level (statistical method) or the 500-mb chart alone.

The basic grid is essentially the same as that used in the 500-mb method except that gradients were computed at intervals of 2.5° lat instead of 5°. The previous 12-hour motion was also incorporated into the forecast.

This method was tested on 23 forecasts during the 1958 hurricane season. The average error was 95 nautical miles for the 24-hour forecast and ranged from 15 nautical miles to 170 nautical miles.

A further explanation of the method and its procedure for application may be found in the Bulletin of the American Meteorological Society, Vol. 41, No. 2, February 1960.

CHAPTER 13

WEATHER BRIEFING AND FLIGHT FORECASTING

As an Aerographer's Mate First Class or Chief, one of your principal duties will be to prepare and give weather briefings of all types. There are many variations of the types of briefings required of you. They range from the individual pilot weather briefing to the more complex and detailed command and staff briefings. Normally, the latter type of briefing would be presented by the Meteorological Officer, but in his absence or by direction, you may be required to perform this type briefing. In the first section of this chapter, general practices, insofar as the varied types of briefings you will be called upon to make, are presented for background information. In later sections of this chapter more specific requirements for briefings are given.

PRINCIPLES OF WEATHER BRIEFING

PURPOSE OF BRIEFING

A weather briefing is the presentation of weather information in a concise, clear, and orderly form to those personnel planning, controlling, or participating in military operations. The briefing, no matter which of the many forms it takes, is conducted to insure that planning, supervising, and operating personnel receive and understand the best available weather information, so as to permit these personnel to properly evaluate and utilize the effect of weather factors on projected operations.

To be of maximum value to using personnel, the first principle of successful weather briefing is to insure that the weather information given is tailored to meet the requirements of those using personnel. The forecaster should take the initiative in ascertaining the requirements of his users. Secondly, to be of maximum utility in planning or carrying out operations, the weather information presented should be evaluated for potential accuracy. For example, weather forecasts for the continental United States can usually be expected to verify with reasonable accuracy, due to the excellent network of observation stations and communications facilities. The accuracy of forecasts for areas where data are sparse and communications unreliable would normally be expected to be somewhat less. Using agencies should be apprised of the limitations which such conditions sometimes impose upon the accuracy of the forecasts in order that plans may be adjusted to minimize the effect of an inaccurate forecast upon such operations. Indicating the degree of confidence in a forecast is usually permissible and should not be considered as "hedging." The intent of such a confidence factor is to provide the using agency with the best information available.

PRESENTATION

A weather briefing may be, in many cases, the basic information which determines whether a mission or a large-scale operation is executed as planned, postponed, altered, or canceled. The manner in which the weather information is presented is considered to be of equal importance with the material contained in the briefing.

The forecaster should understand that the information he presents will assist the using agency in clearly determining the effect of the weather on projected operations. He is the authority on an extremely important element affecting the operation, and therefore he should be confident, convincing, and positive. He must present the briefing so that it is clearly heard and easily understood.

Effective weather briefing is an art. It requires alertness, poise, and judgment on the part of the briefer. Whether the information he presents is in oral, written, or pictorial form, he must be able to think on his feet and come up with precise factual answers to questions which may at times be difficult or even embarrassing. He should avoid verbal ambiguities, vagueness, or misplaced emphasis which could easily convey a mental picture of weather completely different from the one intended.

Some helpful briefing hints are listed below:

1. Strive for force and enthusiasm. Show that you are interested in the problem at hand and its correct solution. Remember that all weather briefings are important.

2. Practice the briefing beforehand, (when possible), from opening statement to conclusion, and insure that it is an orderly presentation.

3. Avoid technical meteorological terms. For example, use the word "high" instead of "anticyclone."

4. Avoid ambiguous and vague phraseology. The terms "about," "probably," "I think," etc., convey no useful information and can easily create an impression of hedging.

5. Make the entire briefing a running narrative, and interpret the weather information in terms applicable to the operation. The briefing forecaster does not make operational decisions, but presents an accurate and complete description of the weather to be encountered.

6. Discuss only the important or essential details of the weather. Keep the briefing as simple as possible.

7. Present the briefing in the second person. The material is being presented for and to the audience and is not a mere recitation of a meteorological situation or a map discussion.

8. Never apologize for any part of the weather presentation. You have presented the best available information. If your confidence in the information is limited, you may state so.

9. Finish forcefully, with a definite closing statement. Always ask for questions, as further clarification may be required.

The hints given are merely restatements of the principles of good public speaking. Of course, not every briefing situation will necessarily employ all of these hints. However, you should remember that no matter what type briefing you are giving, the material you are presenting should be heard and understood.

VISUAL AIDS

Visual aids, as used in weather briefings, are pictorial and graphical representations used to portray the information imparted orally by the briefing forecaster. Just as the oral portion of the briefing must be heard, the visual aids must be seen. The size of the pictorial representation depends upon the farthest distance the person receiving the briefing is seated. Color is a valuable aid. Shades of the primary colors contrast best with shades of another primary color. Light shades of colors should be avoided. You do not necessarily need artistic talent for producing effective briefing charts. If the chart neatly and accurately depicts the information, and the datum is legible, it will be satisfactory. The primary purpose of the visual aid is to enhance understanding. Accuracy should never be sacrificed for artistry.

Materials and Equipment

The drawing medium which gives the greatest brilliance and contrast between colors is preferred. The choice of drawing media is limited by the material on which the illustration is made. Illustrations prepared on small cards for projection on a screen require the use of colored pencils or inks of the conventional type used in map analysis.

The briefer should have a basic familiarity with ordinary projection devices such as the transparent projector, the opaque projector, and/or slide projectors, when appropriate. The briefer should also know the capabilities and limitations of the devices used locally.

Types of Visual Aids

Facilities available to the weather office and the requirements and preferences of the using personnel determine which of the various methods of visualizing the weather may be used most advantageously. In briefing an individual pilot, the station-drawn charts, or facsimile charts received by the station, are normally adequate; for larger briefings, a number of other methods may be employed.

OPAQUE PROJECTION.–Charts and cross sections may be drawn on suitable paper and projected by use of a projector fitted with a mirror and lens arrangement for screening opaque materials. For recurring missions, such paper may be overprinted with the terrain features. Opaque projection requires a fully darkened room in order to show details and coloring satisfactorily.

TRANSPARENT PROJECTION.–Transparent projection may be in the form of slides, either locally constructed or ones prepared by some other source. Transparent overlays with the use of the overhead projector lend themselves more readily to locally prepared visual aids. A transparent overlay of plastic can be mounted over a base map or enlarged cross section blanks. A weather picture drawn on the transparent plastic with a china marking pencil can be erased with a soft cloth. When the number of personnel briefed is small, acetate with a frosted surface may be used to prevent glare. However, marks and shadings put on with oily crayons cannot be removed from the frosted surface easily. For command and staff briefings where preparation time is not of the essence, effective presentation can be made by painting on glass plates. Topography and lines designating height and altitude can be drawn permanently on the glass. If the background is covered with blue, an illusion of depth and a maximum contrast with white clouds is obtained.

ENLARGED MOUNTED CHARTS.–Enlarged, expendable type charts are the most desirable for briefing large groups, if time and materials are available. Advantages of this type chart are that the room need not be darkened, charts can be of any desired size, and colors usually contrast more sharply than on projected or overlay type charts.

THREE DIMENSIONAL CROSS SECTION.–For command and staff briefings, a three-dimensional mount may be prepared. This device usually pictures the surface prognostic chart with parallel cross sections mounted vertically on it to represent the weather in natural perspective.

Whatever method is used for briefings, other than for individual pilots, they should be carefully considered in relation to the type of briefing and the maximum benefit of the using personnel.

Preparation of Charts

The information to be displayed on briefing charts varies with the type of mission. This section should be used as a basic guide, bearing in mind that each situation must be resolved in terms of the information desired, the materials available, and the limitations of the briefing.

PROGNOSTIC CHARTS.–The following should be used as a guide for the briefer in preparing prognostic charts. The area shown should be large enough to represent the general weather picture pertinent to the operation. Boundaries between land and water areas may be shaded lightly in blue on the water side and in brown on the land side. Entire land and water areas may be shaded as desirable. All printing should be in black. The date and time should be indicated in the lower margin and should be in the same time zone as used in the briefing. A solid brown line with arrowheads showing direction should be used to indicate the planned route. For convenience and clarity, the route may be divided into zones. When used, each zone should be numbered. Fronts and precipitation areas should be shaded, using the conventional symbols, a sufficient number of isobars should be drawn to show relative intensity and distribution of pressure systems. They should be drawn as solid black lines. Areas of blowing dust, fog, haze, smoke, and similar restrictions to visibility should be shaded lightly in red. Yellow will not ordinarily register visually. The surface flow pattern may be indicated with blue arrows to denote the movement of cold air and with red arrows to show the movement of warm air. Pressure systems should be labeled in accordance with conventional methods. Cloud amounts, the cloud types, and the heights of bases and tops above MSL may be printed on the chart in pertinent areas. Shading or stippling may be used to indicate cloud areas. Weather hazards such as thunderstorms, showers, turbulence, and icing should be indicated by overprinting the appropriate symbol in the pertinent area. Projected movements of fronts may also be shown. For prognostic charts used in planned briefings, additional details may be added and some of those listed above eliminated.

HORIZONTAL WEATHER DEPICTION CHARTS.–These charts are included whenever the Flight Weather Packet is requested. They are discussed in detail later in the chapter.

OPTIONAL CHARTS.–Occasionally, charts other than those listed previously are necessary to complete the visual aids required to insure a satisfactory briefing. One such chart is a cloud distribution and visibility chart. On these charts certain criteria for the type of mission to be briefed are selected (for example 5/10 cloud for one type of bombing mission is critical), and these areas analyzed and indicated by colors, stippling, shading, and outlining the areas as desirable. Other weather elements may also be indicated. Climatological briefing charts may also be used as aids and guides for constructing similar briefing aids for other areas.

TYPES OF BRIEFINGS

Briefings may encompass a very wide range, both as to content and scope, from briefing an individual pilot for a short flight to more elaborate briefings conducted for a high ranking naval officer and his staff for an operation scheduled many weeks hence. Between the extremes there are as many types of weather briefings as there are varieties of military operations. Listed below are general guidelines for some of the briefing types. More specific information is listed later in the chapter.

Ordinary Weather Station Briefing for Individual Pilots

This is by far the most common type of briefing provided by the average weather forecaster. A duty forecaster may perform this type of briefing from 300 to 900 times a month at the average weather station.

The weather briefing should begin as soon as the pilot states his destination and route. The forecaster should keep up to date on current weather on routes commonly flown from his station, in order that the briefing may be conducted quickly, concisely, and without excessive delay in checking current conditions. Particularly for scheduled flights, this system will prove of value. Existing and forecast weather along the route and at the destination, and alternates if required, should be described orally in order to aid the pilot in making his flight plan. This requires in some degree the use of many, if not all, of the following aids: The latest surface synoptic charts, latest surface prognostic chart, latest upper air analyses at applicable levels along with the progs at these levels, route cross sections which are kept current, teletype data for latest hourly sequences, terminal forecasts, pilot reports, radar reports, and any other information deemed of value. Of course, jet aircraft briefing requires data at higher altitudes plus jetstream and contrail data. Overseas stations frequently are called upon to forecast "D" values.

Group Briefings

Group briefings as used in this section are those briefings given to a number of pilots or aircrews as a group, preparatory to a flight. The flight may be a combat mission, a ferry flight, transport, or training flight. For simplification, group briefings may be broken down into two types–cross-country flights, which include usual ferry, transport or training flights; and operational briefings, which include various types of combat or combat training missions.

CROSS-COUNTRY BRIEFINGS.–In general, the content of such briefings should be as follows:

1. Forecast Weather. A brief discussion of the prognostic chart should open the briefing. This discussion should be simple and brief.

2. Route Weather. After the general weather discussion, weather that will be encountered by the aircrews during the flight should be presented. It should include, as part of the oral briefing, takeoff data such as surface wind, weather, visibility, and variations expected; cloud conditions at base and enroute, also including cloud bases below flight altitude with data on coverage, types, bases, tops turbulence,

and/or icing (usually given in heights above mean sea level); freezing level and icing zones; climb wind (climb wind is the mean vector from the surface to flight altitude); weather conditions along path to include winds and temperatures at flight altitude, air-to-air and air-to-ground visibility; a descending wind if appropriate, cloudiness at terminal, weather at terminal, and such landing data with variations such as duration of local showers or thunderstorms. Alternates should be discussed in the same manner as terminals. Additional data may be given, depending on the type of flight, but should be integrated into the body of the briefing at the proper time interval.

For this type of briefing the minimum visual aids which should be used include an HWD chart and the latest Upper Winds chart. When required, a flight folder may be issued to the crews to include a copy of the HWD chart, Upper Winds chart, and a written forecast with additional information.

OPERATIONAL BRIEFINGS.—For operational missions, certain data are required depending on the type of mission. Any or all of the following may be necessary: winds at flight level and at the surface; mean temperature; visibility, both air-to-air and air-to-ground; altitude data; Q factors; refractive index data; altimeter settings; and sky cover and weather at the destination.

Command and Staff Briefings

Briefings of this type can be divided into two general classifications—short range briefings in which weather considerations are presented as they will affect operations taking place in a few hours or a few days; and long-range briefings in which weather conditions are presented as they will affect operations taking place weeks, months, or even years later.

SHORT-RANGE PLANNING BRIEFINGS.—Briefings for short-range planning are directly applicable to impending military operations, and their content will depend on the type of mission. In general, the following information should normally be included:

1. A review of the past and present synoptic situation with a logical transition to weather expected from prognostic charts, including lows, highs, fronts, general weather in each area, cloud cover, hazards, turbulence, icing and visibilities.
2. Base, route, and target conditions as shown on the HWD chart to include cloud cover, winds, weather, hazards and visibility.

In presenting this type of briefing the synoptic chart can be used to give a review of the present weather; the prognostic chart to present the forecast for an operational area at a particular time; and the HWD chart to show weather along a specific route.

LONG-RANGE PLANNING BRIEFINGS.—Briefings of this type are used in planning amphibious operations and other operations which require long-range planning. In such cases, presentation of climatic data will be required. First, a geographical description of the area should be given followed by interpreting the frequencies and means as they appear on the charts and graphs. Information regarding the limitations and reliability of the climatic data presented should be given in order that the planning staff may give the data proper weight in formulating plans.

In general, visual aids for a long-range briefing can be divided into three classifications:

1. Surface charts, such as storm pattern charts, mean pressure and temperature charts, and windflow charts.
2. Graphs and charts of weather elements which have as their primary purpose the presentation of frequency and means of such weather elements as cloudiness, visibility, and precipitation.
3. Special charts such as ocean currents, state of sea, surf, etc., where applicable.

Mission Weather Indoctrination Briefing

This type of briefing is usually conducted a day or two prior to a scheduled operation. It is purely a weather indoctrination, and no attempt is made to forecast the weather which will prevail over the route or area involved. Typical weather should be depicted with emphasis on the area and frequency of occurrence and average and extreme limits of intensity.

Visual aids used generally include current monthly or seasonal mean charts, such as wind-flow fields, storm tracks, frontal positions, etc.

Special Weather Briefings

Special weather briefings are those conducted for each of the various specialist members of an aircrew or individuals involved in the operation with distinct problems. This type briefing supplements the general briefing. Information of such specialized and technical nature which is of primary interest to each group is presented; for example navigators would desire a detailed discussion of climbing winds, winds and temperatures at flight levels, and descending winds. They may also wish to review surface conditions for information on possible ditching.

DEBRIEFINGS

Normally, weather debriefing consists of interrogation to determine the weather encountered during a mission. Crews which have flown a mission are a logical source of accurate information on actual route and target weather. Data furnished by interrogation may not only provide the only practical means of verifying the forecast, but may be of considerable value in amending the forecast for later flights into the same area.

The forecaster should try to impress the crew member with the value of the weather information they can furnish. However, for the information to be of its utmost value, the exact time, place, and altitude of each observation should be obtained. Weather folders, cross sections, inflight report forms, and checklists help the forecaster to get a complete weather picture.

WEATHER BRIEFINGS FOR SURFACE AND SUBSURFACE OPERATIONS

After obtaining a comprehensive understanding of the operational problems involved, the person who will make the weather briefing is ready for the preparation of his discussion. Since many lives and much costly equipment are generally involved in an operation, preparation cannot be haphazard in nature; it must be exhaustive in compilation of data and detailed in deliberation. Indeed, the success or failure of an operation is often attributable either entirely or in part, to the value of weather information received by the operation commander. The fact that such degree of preparation requires a considerable period of time is obvious and further emphasizes the need for close liaison with the operation commander in order to insure that sufficiently early notification of the requirements for a briefing is received.

WEATHER SYNOPSIS

On many Navy forecasts a weather synopsis is made of the pressure systems and fronts. A general description of the systems and fronts with their location and expected movement is given. Below is an example of a weather synopsis:

The station is under the influence of a cold polar air mass. A cold front passed the station early yesterday afternoon. This cold front is now oriented northeast-southwest along the eastern coast of the United States, extending from a low-pressure area centered over southeastern Connecticut. The front has become stationary through northern Florida and extends westward along the southern coast of Alabama and Louisiana. The warm, moist, tropical air is over-riding the cold polar air mass over this area, causing considerable cloudiness and light intermittent rain. This condition is expected to exist during the next 36 hours. A maritime air mass dominates the entire western part of the United States.

SURFACE SHIP OPERATIONS

Weather briefings for surface ship operations will vary considerably, depending upon the type of ship and the mission at hand. Here again, the meteorologist should be cognizant of all phases of the operation and be able to make comments as to how the various weather elements will influence the unit or units. Long-range weather forecasts are desirable for route planning to avoid bad weather. While it is realized that forecasting beyond 36 hours is difficult, use should be made of the latest long-range forecast procedures and of climatological data for extended periods. A prevailing climatological briefing of personnel (especially the officer in tactical command (OTC) and his staff, the commanding officer, executive officer, operations officer, navigator, OOD's, and other department heads) is recommended for extended cruises. Ship routing is now being done on a large scale with a high degree of success through Optimum Track Ship Routing (OTSR).

One of the concerns for the commanding officer is the danger of encountering a typhoon (or hurricane). You will normally find that warnings from the various weather centrals will keep you fairly well posted, but do not let this lull you into a false sense of security. Make a special chart for keeping track of any storms, and use it in your briefings. You cannot give too much advance notice; remember the ship's speed is limited and advance planning is absolutely necessary.

Sea conditions and wind speeds, as related to the development of seas, are important items for most ships.

Visibility is also important, not only for the safety of the ship, but to determine what speed of advance is necessary to maintain schedules. For example, if an extensive fog bank is expected en route, the commanding officer may desire to increase his speed of advance while the visibility is good in order to reduce the speed for safety considerations while proceeding through the fog. In port, the sailing time may be altered in the event that fog is expected at the scheduled time of departure.

Cloud coverage and ceilings are not normally important; however, in time of war it may be desirable to operate in normally undesirable weather which would provide cover for the ship's movements to aid in avoiding visual detection. The use of frontal areas and squalls for cover was not uncommon in World War II. Bad weather can be of use to someone.

Forecasts of conditions affecting the capabilities of radar and sonar equipment are required of the meteorologist. This is becoming increasingly important, and meteorologists should become acquainted with pertinent facets of oceanography and the sonar qualities of the various ocean currents and should understand how weather conditions affect sonar and radar performance.

Cold weather operations are very demanding on the meteorologist. Here he will find that every bit of information has value, no matter how insignificant it may seem. Sea water temperatures as well as air temperatures are very important, particularly when freezing temperatures may be encountered. An accumulation of ice topside can seriously affect a ship's stability and endanger the lives of all personnel. A coating of ice may increase the chance of visual detection in time of war. Ice forming on the decks of carriers is particularly hazardous, not only delaying operations, but also placing the ship in an unready condition and adding to the ever-present hazards to topside personnel.

Drift ice areas are hazards to navigation and may necessitate additional precautions. Slush ice areas are also hazards, and the meteorologist should know the various ways in which the ship may be endangered. Ice may clog the intake strainers of the sea water chests and thus seriously impair the operation of the ship's engineering department. Ice forming in exposed fireplugs and water breakers may cause them to freeze and burst. Other cold weather effects are extremely probable; hence the meteorologist should realize his responsibilities and use his initiative to point out how the operations of his unit, as well as the fleet, may be affected.

The chill factor which is covered in detail in cold weather manuals should be understood and included in daily briefings. Here the meteorologist will find himself concerned with the protection of topside personnel from the weather elements (wind and temperature) and, in a sense, concerned with how they clothe themselves.

Special evolutions must take place aboard ship upon the occurrence of certain weather. Aerographer's Mates should be familiar with these evolutions, the officers responsible for initiating actions, and the weather conditions under which they will need to happen. An EVOLUTION CHECKLIST should be prepared, listing the weather conditions and the individuals to be notified. Some examples of these evolutions are as follows: (1) Heavy rains–Gunnery Officer; (2) Fog–OOD; (3) Freezing temperatures–Engineering Officer. This evolution checklist should be as complete as possible and therefore requires consultation with many department heads.

Radiological fallout predictions and ballistic wind forecasts should be prepared upon request and at any time when the situation demanding such forecasts arises. Such forecasts are of value to carriers as well as to the other surface ships, and the ships in company not having weather units should receive these forecasts when necessary.

CARRIER OPERATIONS

Each of the various types of aircraft carriers are subject to being affected by almost every known weather element.

Wind force aboard a carrier takes on added importance in that there is such a condition as too little wind. All carriers need about 30 knots across the flight deck to insure safely conducted operations. Greater wind speeds are desired for jet aircraft operations. However, with too much wind across the flight deck, it may become difficult to move aircraft about the deck and conduct launchings and landings. It has been known for a carrier to actually back down in order to conduct flight operations. The flight deck wind is often different than the apparent wind at anemometer level; for this reason, hand anemometer wind observations on the flight deck may be necessitated. Catapult personnel may be able to take the required spot measurements of flight deck winds with the hand anemometer.

Type and duration of precipitation take on added importance, especially at near freezing temperatures. A sheet of ice forming on the deck will make it more hazardous for the flight deck crew;- and with aircraft operating the hazard becomes more acute.

The catapult officer can be aided by knowing in advance of low temperatures in order to adjust his equipment to care for sluggishness due to weather causes. A temperature of 32°F is critical and important aboard ship.

With the advent of the CV as attack platform both for targets above and below the sea surface it is imperative that Aerographer's Mates be cognizant of the interplay of the air-ocean environment. Usually in the role of flagship, the carrier weather office can be expected to produce timely and accurate weather bulletins for not only its air group and self but its small boy escort vessels as well.

Aboard carriers such as LPH's or the newer LHA's special requirements for such information as density altitude and pressure altitude may exist. The many smaller ships, including Light Airborne Multi-Purpose System (LAMPS) equipped vessels and minesweeping forces may develop more special weather requirements with increased helicopter operations. Although minesweepers and destroyers with LAMPS are not technically carriers, they still require consideration as ship-aircraft teams and are significant to the thoughtful forecaster.

AMPHIBIOUS OPERATIONS

Amphibious operations are probably the most demanding type of operation requiring accurate weather information. Every imaginable undertaking (airborne, surface, or subsurface) found in amphibious operations is somehow affected by weather elements. A meteorologist forecasting weather should be fully cognizant of all phases of the operation in order to be as valuable as possible. Meteorological considerations are present in all phases of military planning.

Meteorological considerations usually show that one season, month, or other period is more favorable for the planned D-day than another. In making the evaluation, the following effects should be considered:

1. Effects of wind and pressure upon tidal conditions.
2. Effects of wind, sea, and visibility upon beaching conditions, unloading conditions, sonar conditions, and speed of ships and craft, both during the assault and during the period that beach maintenance continues.
3. Effects of wind, temperature, and humidity on aerosols and smoke.
4. Effects of visibility upon navigation and gunfire.
5. Effects of extreme temperatures on personnel, equipment, and planned operations.
6. Effects of weather conditions upon air operations.

The following considerations may influence the determination of the hour of landing:

1. Wind and sea. Strong winds and heavy seas or swell materially affect the approach of ships, particularly landing craft types to the assault area. Heavy weather may also cause transports and other ships to roll excessively, therefore slowing the operation of hoisting out and loading boats with assault troops and equipment. The effects of heavy weather are particularly intensified at night.

2. Surf. Winds affect not only the state of the sea, but also tides, currents, and surf, and hence, beaching conditions. Dangerous or even impossible swell or surf may also be caused by distant storms when local winds are light. Strong winds and rough seas not only may delay the timing of boat waves, but may produce seasickness among the troops and render many ineffectual prior to reaching the shore. On exposed beaches, the height of the surf may reach such proportions as to make landings impracticable. Other beaches may be somewhat sheltered, but the high surf breaking over outlying bars or shoals may render them unusable.

3. Visibility. Navigation during the approach may be impeded by reduced visibility produced by fog, haze or precipitation; at night their effects are greatly intensified. This may result in delay of the arrival of ships in the initial transport area, where further delays may be suffered in the rendezvous of boats in compliance with the time schedule. Landing ships and craft have difficulty in stationkeeping; under reduced visibility conditions, they may straggle and delay their arrival at the destination and correspondingly delay the execution of attack plans.

4. Flying conditions. Suitability of flying conditions for military operations is determined by a combination of factors, including cloud cover, ceiling, visibility, wind, turbulence, and icing. Air observation in support of the assault waves, pre-H-hour bombing of enemy defenses and beaches, and airborne operations may be restricted or even eliminated by unfavorable flying conditions.

There are many other operational problems that the meteorologist is faced with in amphibious operations, but in the main, the foregoing are his most pressing forecasting responsibilities.

SUBMARINE OPERATIONS

As far as subsurface warfare is concerned, naval oceanography has already developed far enough to play a part in both operational and material activities. Knowledge of the sound conditions often makes it possible to use sonar gear more effectively and is sometimes a factor in deciding the proper tactical deployment of vessels. Knowledge concerning the subsurface distribution of temperature and salinity improves the diving operations of submarines and to a considerable degree affects their choice of offensive and evasive maneuvers.

It is of the greatest importance in submarine operation to be able to change depth efficiently. It often takes an appreciable amount of time to flood or pump the required amount of water. During offensive or defensive operations a delay of a few minutes caused by faulty judgment as to the correct ballast change may be costly. The noise in diving operations is also an important consideration when operating among enemy ships that may be maintaining a listening watch.

Maintaining efficient diving operations would be simple if the buoyance of sea water were everywhere uniform. Since it is not, the variations in density that occur make each dive a separate problem, requiring slightly different tactics. Temperature variations have a big influence on the density of water. With decreasing temperature the density will increase and vice versa. The density of sea water is generally dependent on its temperature, salt content, and the pressure of the surrounding water. Density increases when the salinity or pressure increases, but it decreases when the water expands with increasing temperature. When these properties are known, the density can be determined readily from standard tables.

Other properties of the ocean that a meteorologist should brief the submariners on are the water masses they will encounter in their operations and the effects these masses will have on their operations. They should be briefed on ocean currents and how to make the best use of them. The transmission of sound waves is another consideration that should be discussed.

HELICOPTER OPERATIONS

Helicopter pilots are thoroughly familiar with the flight characteristics of their craft and they determine the effect of meteorological elements on their craft. The elements helicopter pilots are interested in are listed in the following paragraphs.

The wind speed in combination with blade rpm and forward speed can be critical with regard to the rotor blade efficiency and flight safety; therefore, an urgent need exists for accurate wind speed forecasts. Wind direction is important from the navigational aspects of the flight.

Helicopter pilots have a keen interest in density and pressure altitude because it controls the load capacity/operating ceiling of the craft. Weather units aboard ships or stations with helicopters on board should take great pains when computing density, pressure altitudes, and upper air temperatures to facilitate the helicopter pilot's flight planning.

Snow and ice accumulations on the helicopter prevent takeoff completely and must be removed from the craft prior to takeoff. Freezing precipitation and frozen precipitation are extremely dangerous to helicopter flying, and the forecasting of these elements is critical in planning helicopter operations.

Surface temperatures control the type of lubrication for certain systems of the helicopter and are an important element in flight operations involving helicopters.

In addition to the above, there are the normal forecast elements of ceiling, sky, and flight visibility. Under bad weather conditions these elements may well determine the feasibility of helicopter operations, especially at sea, even though most military helicopters are equipped with radar equipment.

FLIGHT FORECASTING

One of the major tasks of Aerographer's Mates First Class, and Chief Aerographer's Mates is forecasting the weather for flight operations. Much of your time is spent in completing Flight Weather Briefing Forms and briefing and debriefing pilots on the weather aspects of their flights. Flight forecasting is a comprehensive, difficult, but very interesting task. The purpose of this section is to help to equip senior Aerographer's Mates with the necessary information to adequately perform flight forecasting duties.

Although pilots are responsible for reviewing and being familiar with weather conditions for the area in which the flight is contemplated, weather briefings shall be conducted by a qualified meteorological forecaster, when available. These briefings may be conducted in person or by telephonic, autographic, or weathervision means.

FLIGHT WEATHER BRIEFING FORM

A DD Form 175-1, such as the one described later in this section, is completed for all flights to be conducted in accordance with instrument flight rules, when military weather services are available. The forecaster completes the form for briefings conducted in person and for autographic briefings. It is the pilot's responsibility to complete the form for telephonic or weathervision briefings. For VFR flights using the DD-175, the following certification on the flight plan may be used in lieu of a completed DD Form 175-1:

> BRIEFING VOID ________________Z. FLIGHT AS PLANNED CAN BE CONDUCTED UNDER VISUAL FLIGHT RULES. VERBAL BRIEFING GIVEN AND HAZARDS EXPLAINED.
>
> ______________________________
>
> (Signature of forecaster)

If the intended VFR flight plan includes airfields having VFR minima higher than the basic 1,000 feet ceiling and 3 statute miles visibility, it is the responsibility of the pilot to advise the weather briefer of these higher minima.

A Flight Weather Packet, such as the one described later in this section which includes a Horizontal Weather Depiction (HWP) chart, may be requested by the pilot. It is normally requested that pilots allow a minimum of 2 hours for preparation of the packet.

NAVY FLIGHT RULES

Due to the density of air traffic over the world today, it is imperative that there be an organized system for controlling aircraft. There must be specific rules and regulations set forth for the safety of aircraft operations.

The following sections on Visual Flight Rules (VFR) and Instrument Flight Rules (IFR), which are based on OpNav Instruction 3710.7(), acquaints the Aerographer's Mate with the flight rules that deal with weather.

VFR

Table 13-1 lists the basic VFR weather minimums.

Additional VFR weather minimums are listed below:

1. Destination must have 3 miles or more visibility at time of departure and be forecast to remain 3 miles or more during the period 1 hour before until 1 hour after ETA.
2. The distance from the surface of the airport to the lowest cloud layer reported as "broken" or "overcast" within control zones must be at least 1,000 feet at point of departure, or, if a more stringent requirement has been established, at or above the VFR minimum prescribed in the aerodrome remarks section of the DOD Flight Information Publication (IFR Supplement). NOTE: A "Control zone" normally refers to a circular area with a radius of 5 miles of the airport, and any extensions necessary to include instrument approach and departure paths. In addition, the weather at destination must be forecast to remain at least 1,000 feet, or at or above the prescribed published VFR minimums for that aerodrome during the period 1 hour before until 1 hour after ETA.
3. Aircraft may be operated on a visual flight rules clearance above "broken clouds" or an "overcast" provided climb to and descent from such "on top" flight can be made in accordance with visual flight rules.
4. Because of their special flight characteristics, helicopters may be flown at less than the minimums established above, subject to applicable FAA regulations. VFR flight minimums for clearance of helicopters are established as 1 mile forward flight visibility and 500 foot ceiling.

Table 13-1.—Basic VFR weather minimums.

Altitude	Flight visibility	Distance from clouds
1,200 feet or less above the surface (regardless of MSL altitude)—		
Within controlled airspace	3 statute miles	500 feet below. 1,000 feet above. 2,000 feet horizontal.
Outside controlled airspace	1 statute mile (except as in 91.105(B))	Clear of clouds.
More than 1,200 feet above the surface but less than 10,000 feet MSL—		
Within controlled airspace	3 statute miles	500 feet below. 1,000 feet above. 2,000 feet horizontal.
Outside controlled airspace	1 statute mile	500 feet below. 1,000 feet above. 2,000 feet horizontal.
More than 1,200 feet above the surface and at or above 10,000 feet MSL	5 statute miles	1,000 feet below. 1,000 feet above. 1 mile horizontal.

IFR

The weather criteria used for IFR flights are as follows:

1. IFR clearance is based on the actual weather at the point of departure at the time of clearance, and forecast weather en route and at both the destination and destination alternate during the period 1 hour before until 1 hour after ETA. Existing weather may be used as basis for clearance when no forecast weather is available and the pilot's analysis of available data indicates satisfactory conditions for the planned flight. No clearance can be authorized for destinations at which the weather is forecast to be below minimums upon arrival, unless an alternate airfield is available which is forecast to be equal to or better than 3,000 feet ceiling and 3 miles visibility during the period 1 hour before until 1 hour after the ETA. Flights must be planned to circumvent areas of forecast atmospheric icing conditions and thunderstorms when practicable.

NOTE: The following is an exception to the preceding rule: When urgent military necessity dictates, Commanding Officers of aircraft carriers, Commanders of Fleet Air Detachments, Marine Aircraft Group Commanders, and seniors in the operational chain of command may approve flight plans for aircraft under their cognizance in weather conditions below the prescribed minima as specified in OpNav Instruction 3710.7( ).

2. Aviation Severe Weather Watch Bulletin (WW). The National Weather Service issues unscheduled WW's whenever there is a high probability of severe weather development. These WW's are for a designated area and a specified time period. The WW's are used by the Naval Weather Service for forecasting hazardous flying conditions. The Air Force issues scheduled Military Weather Warning Advisories (MWWA). These graphical advisories are an estimate of the weather producing potential of the existing air mass. These advisories will be given to all pilots filing from all U.S. Air Force bases and will be used for flight planning when National Weather Service WW information is unavailable. Air Force advisories do not constitute a National Weather Service Weather Watch Bulletin. Except for operational necessity, emergencies, and flights involving all weather research projects or weather reconnaissance, approval authorities shall not approve flights nor shall pilots who possess approval authority file through areas for which the National Weather Service has issued an aviation severe weather watch bulletin (WW) unless one of the following exceptions apply:

a. Storm development has not progressed as forecast for the planned route. In such situations, the following rules apply:

(1) VFR clearance may be permitted if existing and forecast weather for the planned route permits such clearance.

(2) IFR clearance may be permitted if aircraft radar is installed and operative, thus permitting detection and avoidance of isolated thunderstorms.

(3) IFR clearance is permissible in positive control areas if visual meteorological conditions can be maintained, thus enabling aircraft to detect and avoid isolated thunderstorms.

b. Performance characteristics of the aircraft permit an en route flight altitude above existing or developing severe storms.

3. An alternate airfield is not required when the weather at the destination is forecast to be equal to or better than 3,000 feet ceiling and 3 miles visibility during the period 1 hour before until 1 hour after the ETA. Otherwise, an alternate airfield is required.

FLIGHT WEATHER BRIEFING FORM

The Flight Weather Briefing Form, DD Form 175-1, provides a comprehensive record of weather flight planning information for pilots and flight clearance authority. This form should be completed with the utmost diligence and care in accordance with the policies set forth in NavWeaServCom Instruction 3145.1( ); it should be completed by qualified Naval Weather Service personnel as set forth in NavWeaServCom Instruction 3140.5( ).

BRIEFING PERSONNEL

The completion of Flight Weather Briefing Forms and briefings of pilots will be carried out by personnel who possess written authorization to conduct such briefings.

To qualify as a meteorological/oceanographic forecaster a candidate must satisfy one of the following academic requirements:

1. Completion of a course of instruction in meteorology at the USN Postgraduate School.
2. Attainment of a degree in meteorology from an accredited university.
3. Completion of AG"B" School.
4. Qualification as a meteorological forecaster in the USAF or the National Weather Service.

The candidate must also fulfill all of the following general requirements:

1. Demonstrate proficiency in briefing and forecasting to the satisfaction of the Commanding Officer, Officer in Charge, or Chief Petty Officer in Charge of the activity.
2. Be indoctrinated in local meteorological/Oceanographic phenomena (Local Area Forecasters Handbook) and in local operations and procedures. Unless exempted by the Commanding Officer/Officer in Charge, all flight forecasters will participate in familarization flights in their respective local flying areas.
3. Be thoroughly familiar with current International Civil Aviation Organization (ICAO) Regulations, and Navy Directives as they pertain to aircraft operations of the activity

Personnel meeting the above criteria may be be authorized to conduct meteorological/oceanographic briefings for which qualified and to prepare forecasts for aircraft flights. This authorization will be in writing and signed by the Commanding Officer, Officer in Charge, or Chief Petty Officer in Charge of the activity.

WEATHER ENTRIES ON DD FORM 175-1

After the pilot has completed the Flight Clearance Form (DD Form 175), the forecaster completes the weather entries on DD Form 175-1.

Entries

All weather entries should be made in the Airways Code form (hourly sequence) as described in the effective edition of FMH No. 1, Surface Observations. Temperatures should be entered in degrees Celsius (°C), wind direction in tens of degrees, and wind speed in knots. All dates/times should be entered in GMT. Flight levels and heights of all meteorological phenomena should be entered in hundreds of feet MSL except as noted. ALL blocks on the form should be completed as described in NavWeaServCom Instruction 3145.1( ). This instruction also contains instructions for ordering new supplies of forms.

Figure 13-1 is an example of entries on a completed Flight Weather Briefing Form, DD Form 175-1.

FLIGHT WEATHER BRIEFING PACKET

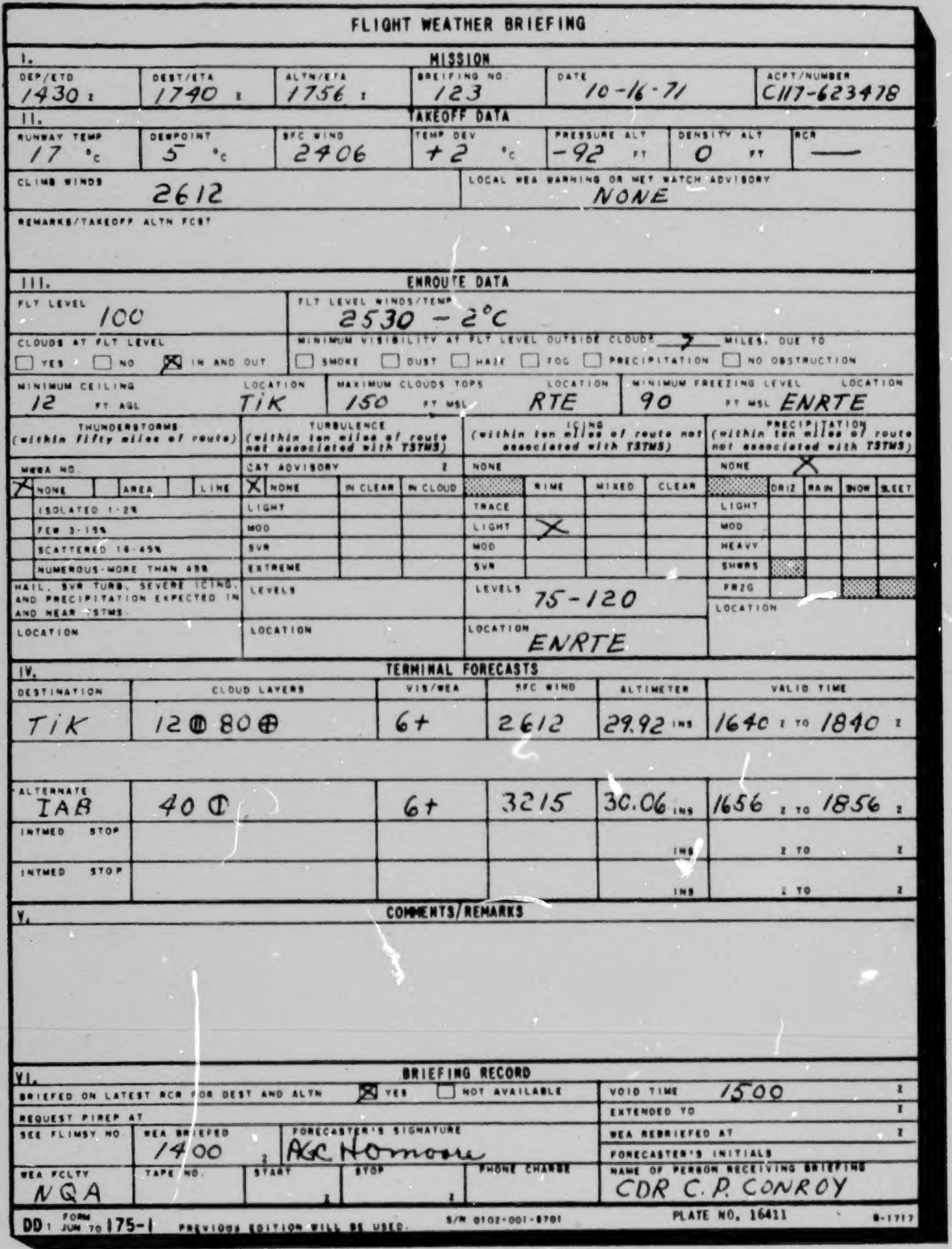
The flight weather packet, which is issued upon the pilot's request, is enclosed in a kneeboard size flight forecast folder for the convenience of the pilot. The reduced size (5" X 8") of the folder allows pilots to clip it to the kneeboard when logging en route weather. The folder has entered on it a variety of weather information as illustrated in figure 13-2.

In addition to the forecast folder, the packet contains the DD Form 175-1, an HWD (high or low level), and upper wind chart(s) for the proposed flight level. When appropriate, a "ditch heading chart" for over water flights and "predicted altimeter setting charts" for over water flights 1,500 feet or below, or when requested, are enclosed.

The charts enclosed should be prepared using standard entries as described on page two of the forecast folder.

FOLDER DESCRIPTION

Although the folder is illustrated in figure 13-2, the following description is provided for clarity:



FLIGHT WEATHER BRIEFING

I. MISSION

DEP/ETD 1430 Z	DEST/ETA 1740 Z	ALTN/ETA 1756 Z	BREIFING NO. 123	DATE 10-16-71	ACFT/NUMBER C117-623478

II. TAKEOFF DATA

RUNWAY TEMP 17 °C	DEWPOINT 5 °C	SFC WIND 2406	TEMP DEV +2 °C	PRESSURE ALT -92 FT	DENSITY ALT 0 FT	RCR —

CLIMB WINDS 2612

LOCAL WEA WARNING OR MET WATCH ADVISORY NONE

REMARKS/TAKEOFF ALTN FCST

III. ENROUTE DATA

FLT LEVEL 100

FLT LEVEL WINDS/TEMP 2530 - 2°C

CLOUDS AT FLT LEVEL ☐ YES ☐ NO ☒ IN AND OUT

MINIMUM VISIBILITY AT FLT LEVEL OUTSIDE CLOUDS 7 MILES, DUE TO ☐ SMOKE ☐ DUST ☐ HAZE ☐ FOG ☐ PRECIPITATION ☐ NO OBSTRUCTION

MINIMUM CEILING 12 FT AGL LOCATION TIK

MAXIMUM CLOUDS TOPS 150 FT MSL LOCATION RTE

MINIMUM FREEZING LEVEL 90 FT MSL LOCATION ENRTE

THUNDERSTORMS (within fifty miles of route)

WWA NO.

☒	NONE	☐ AREA	☐ LINE
☐	ISOLATED 1-2%		
☐	FEW 3-15%		
☐	SCATTERED 16-45%		
☐	NUMEROUS-MORE THAN 45%		

HAIL, SVR TURB, SEVERE ICING, AND PRECIPITATION EXPECTED IN AND NEAR TSTMS.

LOCATION

TURBULENCE (within ten miles of route not associated with TSTMS)

CAT ADVISORY Z

	IN CLEAR	IN CLOUD
☒ NONE		
LIGHT		
MOD		
SVR		
EXTREME		

LEVELS

LOCATION

ICING (within ten miles of route not associated with TSTMS)

NONE

	RIME	MIXED	CLEAR
TRACE			
LIGHT	X		
MOD			
SVR			

LEVELS 75-120

LOCATION ENRTE

PRECIPITATION (within ten miles of route not associated with TSTMS)

NONE X

	DRIZ	RAIN	SNOW	SLEET
LIGHT				
MOD				
HEAVY				
SHWRS	▨			
FRZG			▨	▨

LOCATION

IV. TERMINAL FORECASTS

DESTINATION	CLOUD LAYERS	VIS/WEA	SFC WIND	ALTIMETER	VALID TIME
TIK	12 ⦷ 80 ⊕	6+	2612	29.92 INS	1640 Z TO 1840 Z
ALTERNATE IAB	40 ⦶	6+	3215	30.06 INS	1656 Z TO 1856 Z
INTMED STOP				INS	Z TO Z
INTMED STOP				INS	Z TO Z

V. COMMENTS/REMARKS

VI. BRIEFING RECORD

BRIEFED ON LATEST RCR FOR DEST AND ALTN ☒ YES ☐ NOT AVAILABLE

VOID TIME 1500 Z

REQUEST PIREP AT

EXTENDED TO Z

SEE FLIMSY NO.

WEA BRIEFED 1400 Z

FORECASTER'S SIGNATURE AGC Homoore

WEA REBRIEFED AT Z

FORECASTER'S INITIALS

WEA FCLTY NQA

TAPE NO.

START Z

STOP Z

PHONE CHARGE

NAME OF PERSON RECEIVING BRIEFING CDR C.P. CONROY

DD FORM 1 JUN 70 175-1 PREVIOUS EDITION WILL BE USED. S/N 0102-001-8701 PLATE NO. 16411 0-1717

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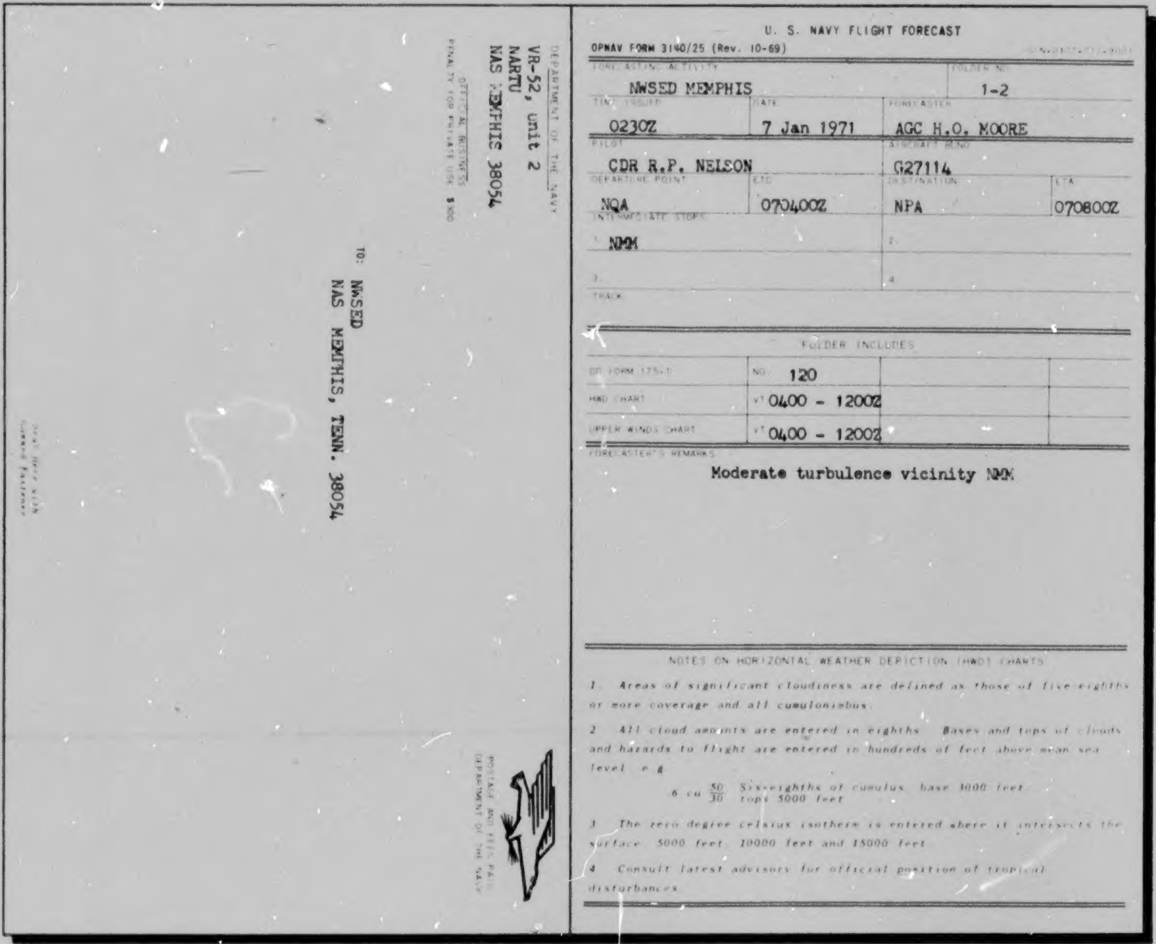
Figure 13-1.—Example of entries on Flight Weather Briefing Form, DD Form 175-1.



DEPARTMENT OF THE NAVY
VR-52, Unit 2
NARTU
NAS MEMPHIS 38054
OFFICIAL BUSINESS
PENALTY FOR PRIVATE USE $300

TO: NWSED
NAS MEMPHIS, TENN. 38054

Seal Here with
Gummed Fastener

POSTAGE AND FEES PAID
DEPARTMENT OF THE NAVY

U. S. NAVY FLIGHT FORECAST

OPNAV FORM 3140/25 (Rev. 10-69)

FORECASTING ACTIVITY: NWSED MEMPHIS
FOLDER NO.: 1-2
TIME ISSUED: 0230Z
DATE: 7 Jan 1971
FORECASTER: AGC H.O. MOORE
PILOT: CDR R.P. NELSON
AIRCRAFT BUNO: G27114
DEPARTURE POINT: NQA
ETD: 070400Z
DESTINATION: NPA
ETA: 070800Z
INTERMEDIATE STOPS: 1. NMM 2. 3. 4.
TRACK:

FOLDER INCLUDES

DD FORM 175-1	NO. 120		
HWD CHART	VT 0400 - 1200Z		
UPPER WINDS CHART	VT 0400 - 1200Z		

FORECASTER'S REMARKS

Moderate turbulence vicinity NMM

NOTES ON HORIZONTAL WEATHER DEPICTION (HWD) CHARTS

1. Areas of significant cloudiness are defined as those of five-eighths or more coverage and all cumulonimbus.

2. All cloud amounts are entered in eighths. Bases and tops of clouds and hazards to flight are entered in hundreds of feet above mean sea level, e.g.

$6\ cu\ \frac{50}{30}$ Six-eighths of cumulus, base 3000 feet, tops 5000 feet

3. The zero degree celsius isotherm is entered where it intersects the surface, 5000 feet, 10000 feet and 15000 feet.

4. Consult latest advisory for official position of tropical disturbances.

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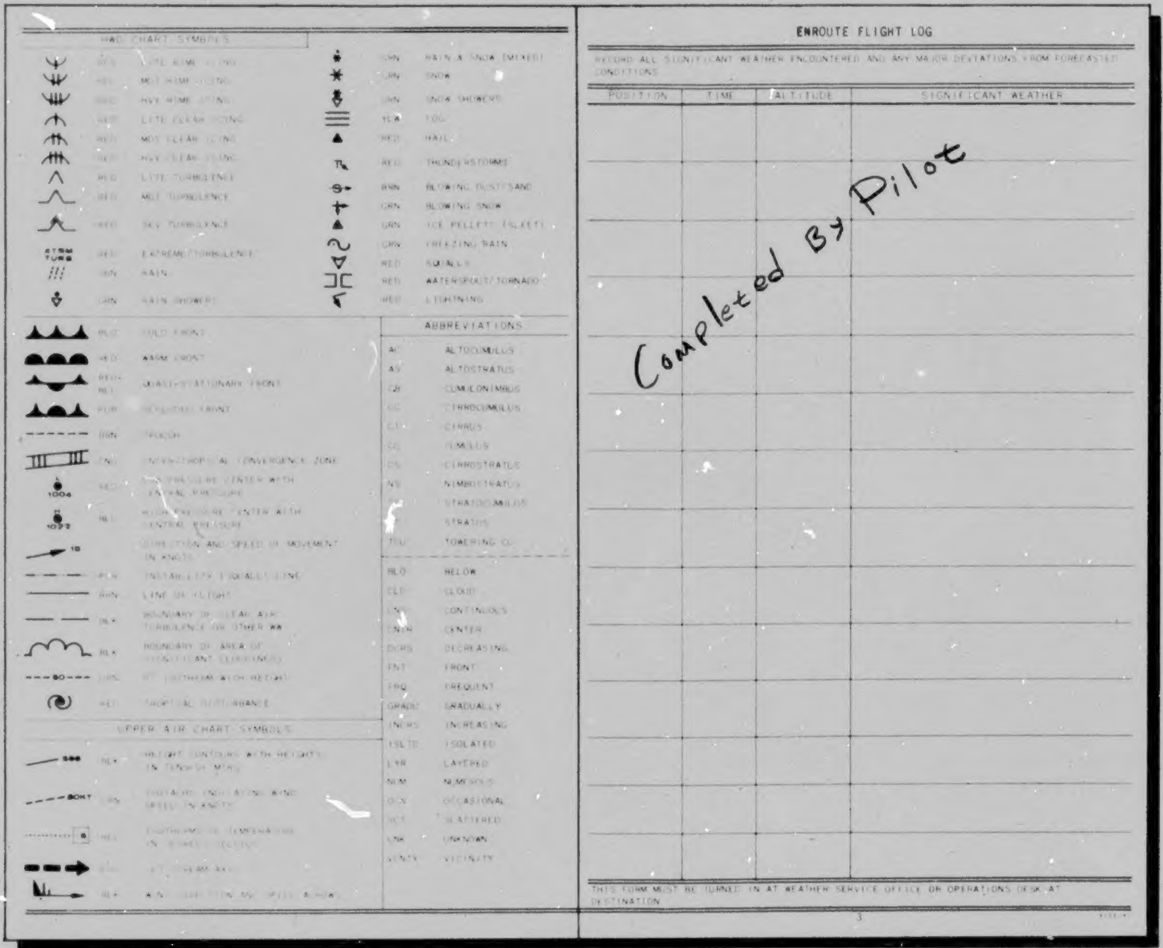
Figure 13-2.—Forecast folder, OpNav Form 3140/25.

1. The section entitled "FOLDER INCLUDES" on page one of the folder lists all forms and charts that must be included in all packets. Other inclusions should be listed in the blank spaces provided.

2. Page two lists the symbols used on the enclosed charts.

3. Page three includes the En Route Flight Log which may be used in lieu of the AIREP FLIGHT LOG, OpNav 3140-30, and should be brought to the attention of the pilots. Aircraft commanders who commonly use the AIREP FLIGHT LOG for transoceanic flights should be encouraged to continue to do so as these AIREPS provide key data for computer analysis centers.

4. Page four is a self-mailer for the folder. The issuing weather activity should enter its address to the right and below the guide dot. The receiving weather activity should review the En Route Flight Log portion, enter its address in the return address section, remove all staples,



HWD CHART SYMBOLS

RED	LITE RIME ICING
RED	MDT RIME ICING
RED	HVY RIME ICING
RED	LITE CLEAR ICING
RED	MDT CLEAR ICING
RED	HVY CLEAR ICING
RED	LITE TURBULENCE
RED	MDT TURBULENCE
RED	SVR TURBULENCE
RED	EXTREME TURBULENCE
GRN	RAIN
GRN	RAIN SHOWERS
GRN	RAIN & SNOW (MIXED)
GRN	SNOW
GRN	SNOW SHOWERS
YLW	FOG
RED	HAIL
RED	THUNDERSTORMS
BRN	BLOWING DUST/SAND
GRN	BLOWING SNOW
GRN	ICE PELLETS (SLEET)
GRN	FREEZING RAIN
RED	SQUALLS
RED	WATERSPOUT/TORNADO
RED	LIGHTNING
BLU	COLD FRONT
RED	WARM FRONT
RED-BLU	QUASI-STATIONARY FRONT
PUR	OCCLUDED FRONT
BRN	TROUGH
ORG	INTERTROPICAL CONVERGENCE ZONE
RED	LOW PRESSURE CENTER WITH CENTRAL PRESSURE
BLU	HIGH PRESSURE CENTER WITH CENTRAL PRESSURE
	DIRECTION AND SPEED OF MOVEMENT IN KNOTS
PUR	INSTABILITY (SQUALL) LINE
BRN	LINE OF FLIGHT
BLK	BOUNDARY OF CLEAR AIR TURBULENCE OR OTHER WX
BLK	BOUNDARY OF AREA OF SIGNIFICANT CLOUDINESS
GRN	0° ISOTHERM WITH HEIGHT
RED	TROPICAL DISTURBANCE

UPPER AIR CHART SYMBOLS

BLK	HEIGHT CONTOURS WITH HEIGHTS IN TENS OF METERS
GRN	ISOTACHS INDICATING WIND SPEED IN KNOTS
RED	ISOTHERMS OF TEMPERATURE IN DEGREES CELSIUS
[illegible]	JET STREAM AXIS
BLK	WIND DIRECTION AND SPEED ARROWS

ABBREVIATIONS

AC	ALTOCUMULUS
AS	ALTOSTRATUS
CB	CUMULONIMBUS
CC	CIRROCUMULUS
CI	CIRRUS
CU	CUMULUS
CS	CIRROSTRATUS
NS	NIMBOSTRATUS
SC	STRATOCUMULUS
ST	STRATUS
TCU	TOWERING CU
BLO	BELOW
CLD	CLOUD
CONT	CONTINUOUS
CNTR	CENTER
DCRS	DECREASING
FNT	FRONT
FRQ	FREQUENT
GRADU	GRADUALLY
INCRS	INCREASING
ISLTD	ISOLATED
LYR	LAYERED
NUM	NUMEROUS
OCNL	OCCASIONAL
SCT	SCATTERED
UNK	UNKNOWN
VCNTY	VICINITY

ENROUTE FLIGHT LOG

RECORD ALL SIGNIFICANT WEATHER ENCOUNTERED AND ANY MAJOR DEVIATIONS FROM FORECASTED CONDITIONS

POSITION	TIME	ALTITUDE	SIGNIFICANT WEATHER

Completed By Pilot

THIS FORM MUST BE TURNED IN AT WEATHER SERVICE OFFICE OR OPERATIONS DESK AT DESTINATION

3

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Figure 13-2.—Forecast folder, OpNav Form 3140/25—Continued.

seal the open side with a gummed fastener, and mail the form within 72 hours of receipt. If the AIREP FLIGHT LOG is used instead of the En Route Flight Log portion, it should be enclosed in the folder.

DESCRIPTION OF INCLUDED CHARTS AND FORMS

The following paragraphs present a brief description of the charts and forms contained in the Flight Weather Packet.

1. DD Form 175-1, Flight Weather Briefing Form. This form is completed as described earlier in this chapter. An original and two copies are completed. Copy two should be retained on file in the weather office for 3 months.

2. Horizontal Weather Depiction Chart (HWD). The HWD chart is basically a strip chart containing operationally significant weather data. This chart is required in all Flight Weather Packets. The basic chart or charts to be used

may be prescribed by individual activities. Whenever possible, base charts, listed in Section 4 of the Department of Defense "Catalog of Weather Plotting Charts," NA-50-1G-518, should be used.

The size of the chart used for the HWD chart included in the Flight Weather Packet should be no wider than 8 inches and no longer than necessary to depict the flight path. The overall size of the chart should be kept to a minimum consistent with legibility of entries to facilitate ease of handling within the cockpit. Chart scales from 1:10,000,000 to 1:20,000,000 are ideally suited to this purpose.

HWD's are discussed in more detail later in this chapter under Flight Forecasts.

3. Upper Wind chart. The Upper Wind chart, valid time nearest midtime of the flight, at or near flight level should be included in the packet. The proposed line of flight should also be entered on this chart.

4. Ditch Heading chart. This chart, consisting of plotted ditch heading arrows or point values, should be included for all over water flights.

5. Predicted Altimeter Setting chart. This chart contains a plot of point values of predicted altimeter settings and should be included for all over water flights 1,500 feet or below, or when requested.

6. Miscellaneous charts. Any other operationally necessary charts for specific flights should be included; e.g., Constant Pressure charts, Streamline charts, Surface Wind charts, Sea Surface Temperature charts, etc.

FLIGHT FORECASTS

Flight forecasts consist of the oral, written and pictorial statements to the pilot relating the expected weather along his flight route and at his destination. Flight forecasts include horizontal weather depiction charts (HWD); route and terminal forecasts; pressure pattern flight (for flight planning); icing; contrails; turbulence; cloud cover; ceiling; weather; visibility; and winds, all incorporated into a flight briefing for the type of aircraft flown and the type of operation planned. The following sections of this chapter provide the Aerographer's Mate with the necessary information procedures, and techniques to prepare an adequate flight briefing for aviators.

HORIZONTAL WEATHER DEPICTION (HWD) CHARTS

The HWD is a prog chart for a fixed valid time as near as possible to mid-time of the flight, and portrays all forecast hazards to flight, related weather patterns, and significant pressure and frontal systems. However, when prepared from available facsimile/NEDN charts, they will verify at the 6-hrly synoptic time nearest the mid-time of the flight. Two basic categories of HWD charts will be issued, depending on the proposed flight level. The low-level HWD covers the stratum surface to 400 mb; the high level HWD covers the stratum 400 to 150 mb.

Low Level HWD

The following steps outline the procedure for preparing a low level HWD:

1. Show boundaries of weather areas of significant cloudiness (5/8's or more cloudiness, including cirriform which would prevent a celestial fix) and all cumulonimbus areas by a black scalloped line.

2. Enter cloud amounts in OKTAS, cloud type; and height of bases and tops in hundreds of feet above MSL, within the appropriate areas. Use standard abbreviations for cloud type. Enter the height of the cloud top above the height of the base and separate with a horizontal line. When more than one layer is forecast to occur, enter the higher layer directly above the entry for the lower layer. Some sample entries are:

$$7\ \text{AS}\ \frac{190}{160}$$

$$6\ \text{CB}\ \frac{200}{30}\ \text{OCNL TOPS TO 230}$$

3. Outline areas of clear air turbulence (CAT) or other WW with a heavy dashed black line. Within these areas indicate the degree of CAT with appropriate symbol followed immediately

by the height of the base and top of the CAT stratum in hundreds of feet and/or appropriate symbol for other WW.

4. Depict frontal positions using standard analysis symbols and colors and pressure centers with large "H" or "L" in standard colors with central pressure in millibars entered below. Indicate direction and speed of movement using vectors for direction with speed in knots at the end of the vector.

5. Indicate the 0° C isotherm with its height in increments of 5000 feet above sea level as a dashed green line.

6. Indicate areas of significant weather, obstructions to vision, and icing, with appropriate symbols followed by heights of the base and top of the phenomena in hundreds of feet.

7. Enter the flight route as a solid brown line.

8. Place an identifying legend in the lower margin of the chart.

EXAMPLE: LOW LEVEL HORIZONTAL
WEATHER DEPICTION
Issued by: NWSED Memphis, Tn.
VT 0600Z 26 April 1973

"VT" is the valid date and time of the chart. "Issued by" is the weather office at which the chart was prepared.

High Level HWD

The following steps outline the procedures for preparing a high level HWD:

1. Using the same techniques as for the low level HWD, depict all areas of 5/8's or more cirriform cloudiness at or above flight level; all areas of convective clouds extending to the proposed flight level; all areas of icing or turbulence including CAT; all areas of severe weather published by Aviation Severe Weather Warnings (WW) or other advisory; and surface fronts and pressure centers (with direction and speed of movement).

2. Enter the flight route as solid brown line.

3. Place an identifying legend in the lower margin of the chart in the same manner as the low level HWD.

When local policies or regulations require operational data additional to that which is contained on the Form DD175-1, a section containing appropriate blocks for displaying the information may be appended to the HWD.

Figure 13-3 is an example of a typical Horizontal Weather Depiction (HWD) chart.

UPPER LEVEL PROGNOSIS

The upper level prognoses are prepared for a standard isobaric surface nearest the flight level with a fixed valid time as near as possible to the mid time of the flight. This chart supplements the horizontal weather depiction chart.

The following steps outline the procedures for preparing an upper level prognosis chart:

1. Draw contours for every 60 meters up to the 300 mb level and for every 120 meters for the 300 mb level and above in solid black lines. Use dashed intermediate contours when needed to clarify the flow. Double the interval in case of a very tight gradient. Label all contours at loose ends and above or below closed centers with height in tens of meters. Contours may also be labeled at any other position also as required to make the chart more easily readable.

2. Draw and label isotachs for every 20 knots.

3. Draw isotherms for the level in increments of 5° C using short dashed red lines. Enclose label-value in a square $\boxed{-5}$.

4. Show jet stream axis as a heavy dashed purple line with an arrowhead to show direction.

5. Indicate flight track as solid brown line.

6. Place an identifying legend in the lower margin of the chart similar to the one used for the HWD chart.

ROUTE AND TERMINAL FORECASTS

Of all the sources of weather information available to the weather office in the continental United States, Alaska, and Hawaii, undoubtedly one of the most valuable is the teletype report. It consists of sequence reports, regional synopses, area forecasts, winds aloft forecasts, and route and terminal forecasts.

National Weather Service Weather Service Forecast Offices (WSFO) issue specific forecasts

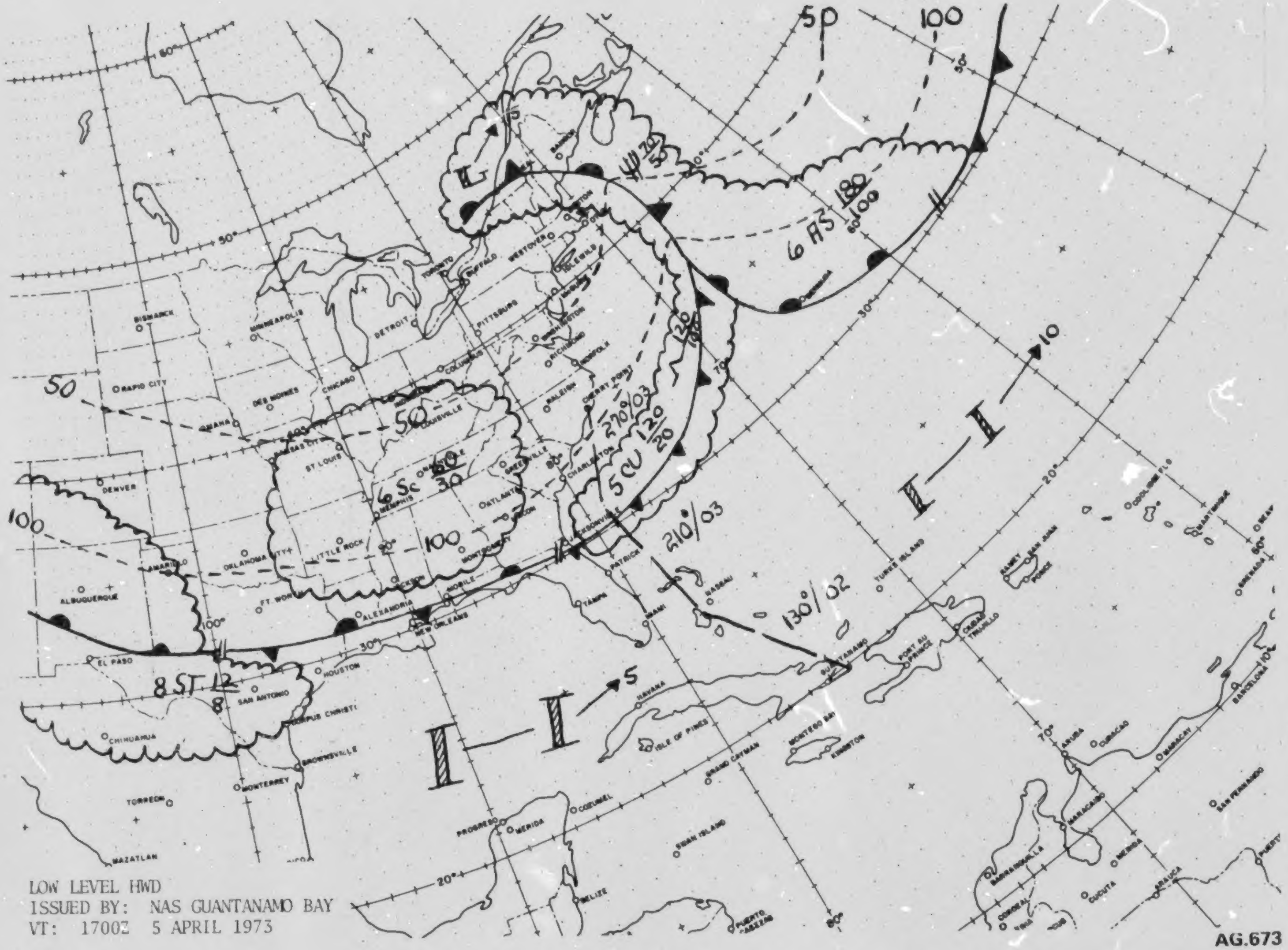
LOW LEVEL HWD
ISSUED BY: NAS GUANTANAMO BAY
VT: 1700Z 5 APRIL 1973
AG.673
50
100
6 AS 180/100
40 20/30
5 CU 120/20
270/03
120/100
210/03
130/02
6 Sc 60/30
50
100
8 ST 12/8
10
5
BISMARCK
RAPID CITY
DENVER
ALBUQUERQUE
EL PASO
AMARILLO
OKLAHOMA CITY
LITTLE ROCK
FT WORTH
SAN ANTONIO
HOUSTON
CORPUS CHRISTI
BROWNSVILLE
MONTERREY
CHIHUAHUA
TORREON
MAZATLAN
MINNEAPOLIS
DES MOINES
OMAHA
CHICAGO
DETROIT
ST LOUIS
MEMPHIS
ALEXANDRIA
NEW ORLEANS
MOBILE
ATLANTA
MONTGOMERY
COLUMBUS
PITTSBURG
TORONTO
BUFFALO
WESTOVER
WASHINGTON
RICHMOND
NORFOLK
RALEIGH
CHERRY POINT
CHARLESTON
GREENVILLE
JACKSONVILLE
TAMPA
PATRICK
MIAMI
NASSAU
HAVANA
ISLE OF PINES
GUANTANAMO
PORT AU PRINCE
TRUJILLO
TURKS ISLAND
SAN JUAN
KINGSTON
MONTEGO BAY
GRAND CAYMAN
COZUMEL
MERIDA
PROGRESO
SWAN ISLAND
BELIZE
PUERTO CABEZAS
ARUBA
CURACAO
MARACAIBO
BARRANQUILLA
CUCUTA
MERIDA
ARAUCA
MARACAY
BARCELONA
SAN FERNANDO
CORO
MARTINIQUE
GRENADA
COOLIDGE FLD
BANGOR
80°
50°
30°
20°
100°
90°
70°

Figure 13-3.—Horizontal Weather Depiction Chart.

for responsible areas at prescribed times throughout a 24-hour period, and these forecasts are transmitted over the teletype in code, using many of the sequence report symbols, abbreviations, and contractions.

For complete information regarding the formats used and times of issue, refer to the Flight Services Manual ATS 7110.10 part II.

Outside the continental United States, Alaska, and Hawaii, these services are provided on a much more limited basis and often are not available at all. In those cases, Aerographer's Mates are required to formulate their own route and terminal forecasts.

Route Forecasts

A pilot must know the weather conditions that now exist and the forecast conditions for his flight route in order to prepare an efficient and safe flight plan. The information he receives from the weather office will help him to make an efficient flight.

The pilot and duty forecaster should study various data and discuss the weather. The forecaster must be sure the pilot thoroughly understands the weather which he will encounter. The forecaster and pilot should study and discuss the surface synoptic charts, noting any low-pressure areas and fronts that are in the general area of the flight route. Possibility of fog and low stratus must be considered. Hourly sequence reports must be studied. The winds aloft and constant pressure charts should be used in determining a satisfactory flight altitude. Various levels should be taken into consideration, since the desired altitude may not be granted by Air Traffic Control. Upper air sounding charts should be studied to determine the temperature at various altitudes, thickness of cloud layers, icing conditions, and stability of air masses.

All available forecasts for the flight route and destination should be read.

The route forecasts are issued by the various regional Weather Service Forecast Offices (WSFO) for the various airways in the United States. Ordinarily, these route forecasts are of the highest quality; nevertheless they may be questioned. In the light of later information or developments, it may become necessary to augment or alter them on occasion. Outside the United States you will often be required to make your own forecast for the whole route.

Terminal Forecasts

Terminal forecasts for most airports in the United States are issued several times daily and revised in the light of further development by the regional Weather Service Forecast Offices (WSFO) and disseminated over the Service A teletype network for use in flight forecasting.

In the United States, then, a flight forecaster's workload is greatly reduced due to the availability of terminal forecasts. Terminal forecasts are also available at military stations on COMET II; TAF or PLATF code form is used.

Outside the United States, terminal forecasts may not readily be available although the meteorological services of most nations do promulgate terminal forecasts in the TAF code form. It is aboard ship that terminal forecasts are least likely to be available, and it is then that the Aerographer's Mate must formulate his own forecast. Helpful details for formulating terminal forecasts may be found in chapter 10 of this manual and Air Weather Service Manual 105-51/1 Terminal Forecasting.

Although the terminal forecasts provided by the various meteorological services are a great help to you at all times, it is inadvisable to accept all the terminal forecasts at face value. Scrutinize the forecasts and make such revisions to them as changing synoptic features may dictate.

PRESSURE PATTERN FLIGHT

Soon after World War II, a system of flight planning, known as pressure pattern flying, was popular as a means of effecting minimal flight time for long distances. For this system, meteorological personnel had to compute and forecast D-values, and then construct D-value charts. Today, it is exceedingly rare for an Aerographer's Mate to be required to construct a D-value chart, but it is quite common for him to have to provide D-values for specific points or areas. Consequently, only determination of D-values will be discussed in this section.

Computing D-values

A D-value describes the height of a pressure surface by its departure from "standard" height.

$$D = Z - Z_p$$

Where: Z is the standard atmosphere altitude above mean sea level.

Z_p is the actual altitude (pressure altitude from a radiosonde sounding) of the same point in space.

For example: The height of the 700 mb level is 9880 feet in the Standard Atmosphere. If the actual height of the 700 mb level at a certain point is 10,200 feet, the D-value is

$$D = 9880 - 10{,}200 = -320 \text{ feet}$$

D-values are of significance for aircraft flying with the altimeter set at 29.92. In the above example, the altimeter of an aircraft flying at the 700 mb level would read 9880 feet, but its actual altitude would be 10,200 feet above sea level. Over water, its radar altimeter would read the latter value.

Forecasting D-values

Forecasting consists of forecasting the actual height of the given pressure surface and then computing the D-value as shown above.

FLIGHT LEVEL WINDS

Forecasting flight level winds or winds aloft in general is an integral part of flight forecasting. Within the continental United States, Alaska, and Hawaii, the task is somewhat simplified in that the National Meteorological Center regularly transmits prognostic constant pressure charts and winds aloft charts on the facsimile system. The facsimile transmissions are further supplemented by winds aloft forecasts sent out by the various regional Weather Service Forecast Offices (WSFO) on the Service C teletype system. These charts and forecasts may be used by flight briefing personnel without further modification under most circumstances, though it is a good idea to constantly revise these forecasts in view of developments occurring sometime in the verifying period. Outside the continental United States, these services may not be readily available, and it is the purpose of this section to furnish Aerographer's Mates with the knowledge necessary to successfully forecast flight level winds.

The basic approach to forecasting winds aloft is to prognosticate the constant pressure surface nearest to the desired level and extrapolate upward or downward. It is permissible to localize the forecast area so long as sight is not lost of large scale developments which might affect the forecast point.

Procedures for progging constant pressure surfaces are given in chapter 8 of this manual. The forecast wind can be picked off the prog chart by simply reading the wind direction, measuring the gradient, and obtaining the speed.

Isotherm-contour relationships are of value when forecasting upper winds in that they may simplify the task a great deal under certain circumstances. When the forecast wind direction is easy to obtain, the speed may be estimated by relating isotherms and contours. Increasing wind speeds occur whenever the thermal gradient increases, as the wind moves the isotherms closer together and decreasing wind speeds can be safely forecast when the direction and speed of the wind and orientation of the isotherms is such as to spread the isotherms further apart as illustrated in figures 13-4 and 13-5 respectively.

When isotherms and contours are in phase and parallel, little change in speeds and direction need be anticipated in a short-range forecast.

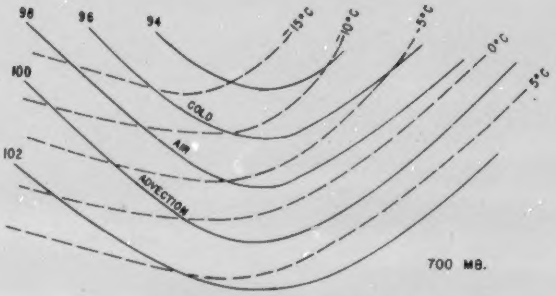

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Figure 13-4.—Increasing wind speed pattern.

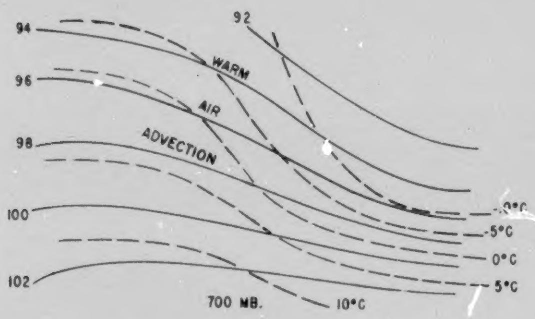

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Figure 13-5.–Decreasing wind speed pattern.

When isotherms and contours are in phase, but not parallel, or out of phase, the situation should be studied more closely, and the prog should be carried out.

FLIGHT WEATHER BRIEFINGS

Procedures for preparing HWD charts; for making route and terminal forecasts; for determining pressure pattern flight routes; and for forecasting icing, contrails, turbulence, and flight level winds have been discussed in the foregoing sections of this chapter.

Methods for forecasting ceilings, weather, and visibility are discussed in chapters 10 and 11 of this manual.

It remains now to integrate the results of the procedures into a flight briefing and to adapt the briefing to the type of operations planned.

CONTINENTAL FLIGHTS

It is the responsibility of the flight forecaster to give the pilot a complete description of existing and expected weather along the route and at the terminals of interest. Upon the flight forecaster's interpretation of the weather will partially depend the type of clearance filed by the pilot. Meticulous consideration of the latest charts and data are necessary for making a reliable forecast for cross-country flights. Airway reports alone are not to be used as the basis for clearance. In briefing, definite statements must be made regarding cloud bases and tops, freezing level, icing zones, turbulence, winds aloft, and frontal weather to be encountered, and possible contingencies resulting from variation of expected trends.

Since each flight presents its own peculiar problems in regard to planning and weather briefing, it is not possible to establish a standard procedure for briefing that will be complete and adaptable to every case. However, experience has shown that there is usually a definite sequence of thought that should be followed when briefing a pilot. This sequence can be logically divided into the following five general steps:

1. Give a brief description of the current weather conditions associated with the line of flight in terms of ceilings, visibilities, cloud layers, winds, turbulence, and icing, as may be appropriate. A brief discussion of the synoptic situation as related to the flight should be included.
2. Give a brief statement of the latest selected observations along and near the flight route.
3. Give a brief statement of the forecast trends of en route weather, identifying significant changes as to expected time of occurrence, and emphasizing areas in which severe weather warnings are in effect or forecast.
4. Give a statement of the expected changes at the terminal and alternate, relating significant forecast changes to the expected time of occurrence.
5. Point out any alternate routes, as appropriate, and discuss weather contingencies. This is the time to volunteer pertinent weather information not included in the preceding discussion and to answer any specific question the pilot may have.

If the pilot requested an HWD chart, present it at this time and discuss the contents with the pilot.

Last but not least, refer the pilot to the various charts on display and discuss them with him.

TRANSOCEANIC FLIGHTS

Transoceanic flights normally operate with limited fuel loads, and a point of no return is computed by the pilot based on the wind forecast. Therefore, it is essential that all pertinent weather data be available to the pilot for advance planning.

The flight packet that is prepared for transoceanic flights is similar to the one discussed for continental flights except that a ditch heading chart must be included for the transoceanic flight and if requested a predicted altimeter setting chart.

The problem of obtaining foreknowledge of pending flights is important. Inasmuch as the thought and work which go into a transoceanic flight forecast must be thorough, flight forecasters should have a minimum of 2 hours of advance notice, although an 8-hour notice is desirable in order to have the material ready for briefing. This problem is solved by close liaison between meteorological personnel and pilots.

When the forecast is complete, it is organized in folder form to be presented to the pilot and fully discussed during the briefing. The number and size of charts should be held to a minimum to save space, and the material should be organized in a neat and readily usable manner.

CARRIER AIRCRAFT BRIEFINGS

Flights of carrier-based aircraft are normally classified as operational and, as such, have flexible weather minimums which are decided upon by the officer in tactical command (OTC). The deciding factor for conducting flights in the event of unfavorable weather conditions will normally be the importance of the mission. The weather briefing of the OTC and/or the commanding officer of the carrier quite frequently will be of considerable importance in deciding whether the scheduled flight operations will be conducted. However, do not forget the pilots.

While it is obviously difficult to conduct briefing of the pilots for all flights when operations are being conducted around the clock, by making maximum use of readyroom briefings, adequate briefings may be made. Most aircraft carriers now have weathervision systems with hookups in the readyrooms, and excellent results are being obtained through maximum and efficient use of this system.

Aerographer's Mates serving as flight forecasters must know the type of operations in which his ship is participating in order to stress the more important elements in his forecast. Weather information for fighter aircraft differs from that desired for ASW flights in that fighter aircraft normally fly at higher altitudes while low-level and sea conditions are more important to ASW operations.

Normally a weather forecast for a flight is for a short period of time, while the planning and scheduling cover a much longer period. Aerographer's Mates will find that their biggest problem is forecasting 24 to 48 hours ahead so that scheduling may be completed. A carrier can move to an area of better weather while a shore base has to sit tight and wait it out. Depending upon the flexibility of the operating area, it may be possible to conduct operations almost continually by shifting to an area where the desired flight operations are feasible.

Surface and upper winds are very important elements for carrier aircraft. This is especially true for fighter aircraft with only the pilot aboard who may find himself too busy to take note of the winds to check his drift. Normally, the CIC of a carrier tracks all aircraft with radar and gives bearings and distances to the aircraft via radio. However, it is possible for the ship to lose contact with the aircraft. The aircraft then is on its own and may have to rely on forecast winds to navigate back to the ship.

Visibility may be important only for launching and landing aircraft; however in the case of fighters or dive bombers, ceiling and visibility in the target area are important. Radar has reduced the visibility requirement for navigation to some degree. For visual searches, it is still important.

Horizontal Weather Depiction charts for use in flight are not normally practical for carrier aircraft.

CHAPTER 14

SEA SURFACE FORECASTING

The task of forecasting the various elements concerned with the sea surface, such as sea waves, swell waves, surf, and surface currents will be encountered by senior Aerographer's Mates filling a variety of billets.

Aboard carriers sea condition forecasts for flight operations, refueling, or underway replenishments will be provided on a routine basis. Staffs of larger facilities will generally provide surf forecasts for amphibious operations, while at air stations supporting search and rescue (SAR) units, forecasts of sea conditions and surface currents may be required at times.

It is therefore important that personnel be familiar with these elements and be able to provide forecasts as necessary.

In this chapter we will discuss these elements as well as methods that may be utilized to forecast the extent of them.

SEA SURFACE CHARACTERISTICS

To accurately forecast sea conditions it is necessary to understand the process whereby waves are developed, the action that takes place as the energy moves, and to have a thorough understanding of the various properties of waves.

This section of the chapter will discuss this background information and terminology used. A complete understanding of these items is necessary to produce the most usable and accurate sea condition forecast.

BASIC PRINCIPLES OF OCEAN WAVES

Ocean waves are advancing crests and troughs of water propagated by the force of the wind. When a wind starts to blow, the sea surface instantaneously becomes covered with tiny ripples which form more or less regular arcs of long radii. As the wind continues to blow, the ripples increase in height and become waves.

A wave is visible evidence of energy moving through a medium, in this case the water, in an undulating motion. As the energy moves through the water, there is little mass motion of the water in the direction of travel of the wave. This can best be illustrated by tying one end of a rope to a pole or other stationary object. When the free end of the rope is whipped in an up and down motion, a series of waves move along the rope toward the stationary end. There is no mass motion of the rope toward the stationary end, only the energy traveling through the medium, in this case the rope.

A SINE WAVE is a true rhythmic progression. The curve along the centerline can be inverted and superimposed upon the curve below the centerline. The amplitude of the crest is equal to the amplitude of the trough, and the height is twice that of the amplitude. Sine waves are a theoretical concept seldom observed in reality. They are used primarily in theoretical groundwork so that other properties of sine waves may be applied to other types of waves such as ocean waves. Principles of other types of waves are modified according to the extent of deviation of their properties from those of sine waves.

Waves which have been created by the local wind are known as SEA WAVES. These waves are still under the influence of the local wind and are still in the generating area. They are composed of an infinite number of sine waves superimposed on each other, and for this reason they have a large spectrum, or range of frequencies.

Sea waves are very irregular in appearance. This irregularity applies to almost all their properties. The reason for this is twofold: First, the wind in the generating area (fetch) is irregular both in direction and speed; second, the many different frequencies of waves generated have different speeds. Figure 14-1 is a typical illustration of sea waves. The waves found in this aerial photograph are irregular in direction, wave length, and speed.

As the waves leave the generating area (fetch) and no longer come under the influence of the generating winds they become SWELL WAVES. For the reason that swell waves are no longer receiving energy from the wind, their spectrum of frequencies is necessarily smaller than that of sea waves. They are smoother and more regular in appearance than sea waves. Figure 14-2 illustrates typical swell wave conditions.

PROPERTIES OF WAVES

All waves have the following properties in common:

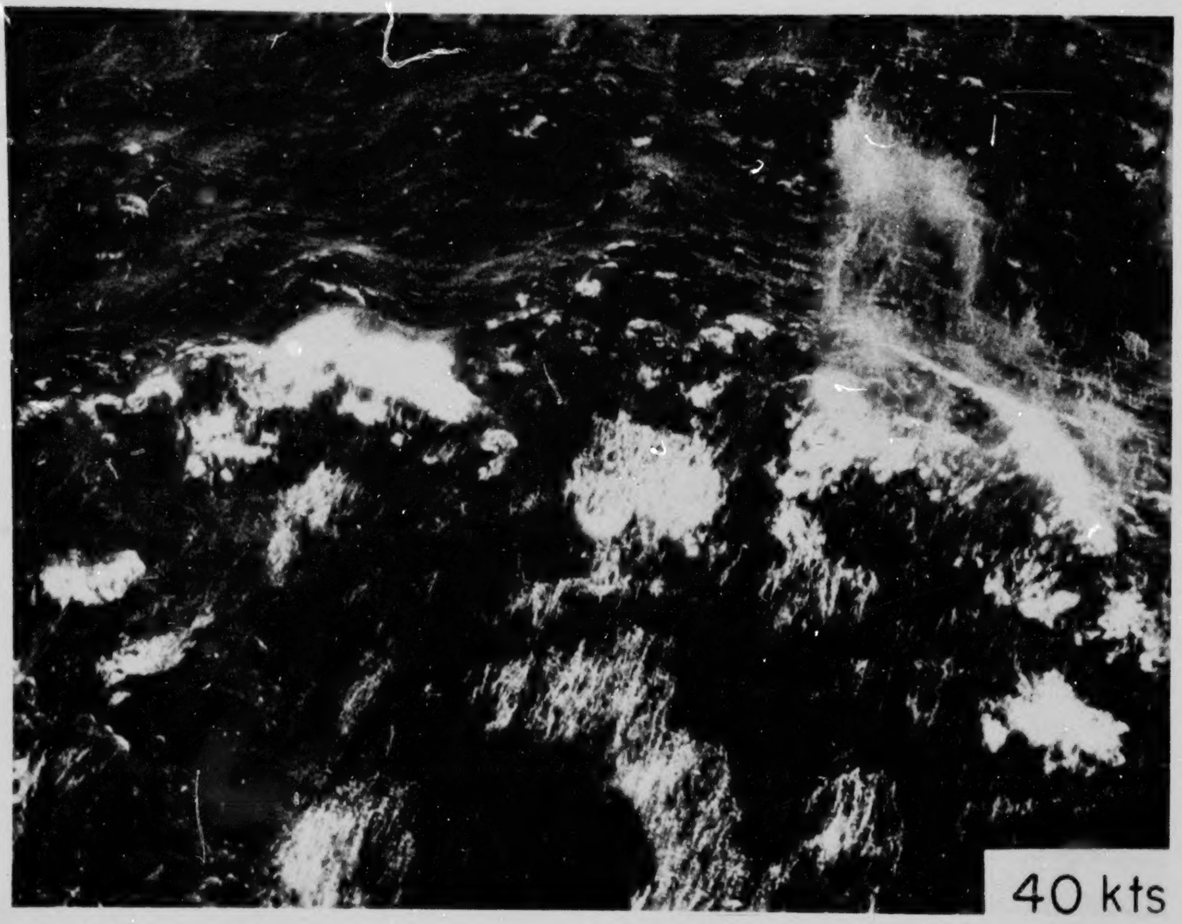

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Figure 14-1.—Aerial photo of sea waves.

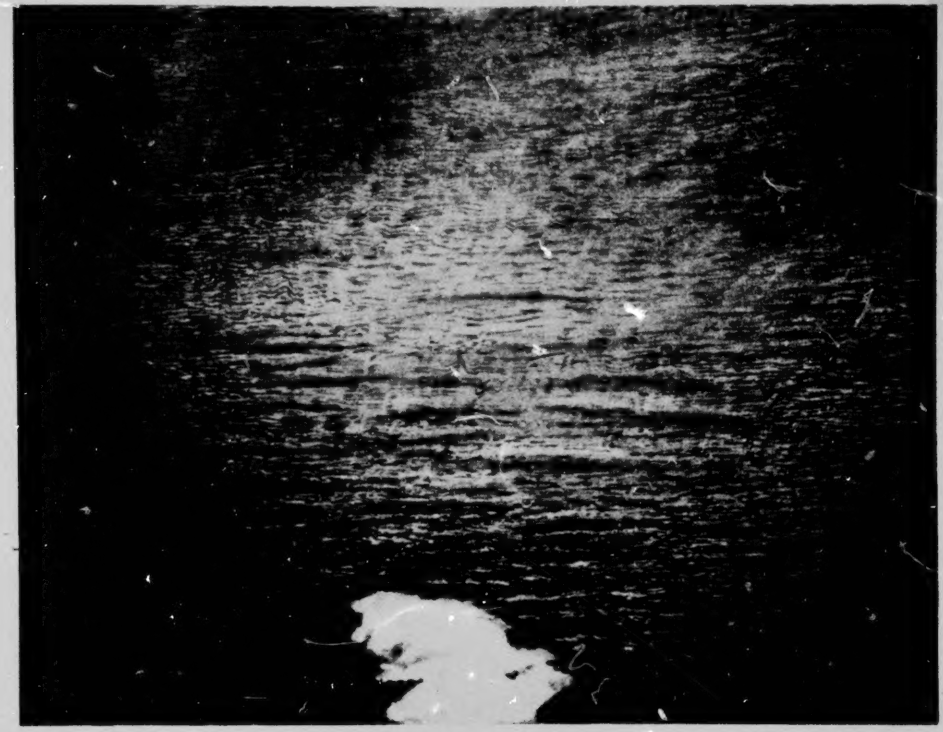

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Figure 14-2.—Swell waves.

1. Amplitude. The amplitude of a wave is the maximum vertical displacement of a particle of the wave from its rest position. In the case of ocean waves, the rest position is sea level.

2. Wave Height (H). Wave height is the vertical distance from the top of the crest to the bottom of the trough. Wave height is measured in feet. Three values for wave height are determined and forecast. They are:

a. H_{avg} (the average height of the waves). This average includes all the waves from the smallest ripple to the largest wave.

b. H_{sig} (the average height of the highest one-third of all waves). The significant height of waves seems to represent the wave heights better than the other values, and it will be used most often for this reason.

c. $H_{1/10}$ (the average height of the highest one-tenth of all waves). $H_{1/10}$ is used to indicate the extreme roughness of the sea.

3. Period (T). The period of a wave is the time interval between successive wave crests, and it is measured in seconds.

4. Frequency (f). The frequency of waves is the number of waves passing a given point during 1 second. It is the reciprocal of the period. In general, the lower the frequency, the higher the wave; the larger the frequency, the smaller the wave. This can be seen when you consider that when at a given point the frequency is 0.05, considerably fewer waves (therefore larger waves) are passing the point than if the frequency were 0.15.

5. Wave Length (L). The wave length is the horizontal distance between two successive crests or from a point on one wave to the corresponding point on the succeeding wave. Wave length is measured in feet, and it is found by the formula: $L = 5.12T^2$.

6. Wave Speed (C). The wave speed is the rate with which a particular phase of motion moves along through the medium. It is the rate at which a wave crest moves through the water. There are two speeds used in ocean wave forecasting: group speed and individual speed. The group speed of waves is approximately one-half that of the individual speed. The individual wave speed in knots is found by the formula C = 3.03T. The group wave speed is found by the formula C = 1.515T.

Definitions of Other Terms

Other definitions with which the Aerographer's Mate should be familiar are as follows:

1. Wave Spectrum. The wave spectrum is the term which describes mathematically the distribution of wave energy with frequency and direction. The spectrum consists of a range of frequencies.

2. Deep Water. Water that is greater in depth than one-half the wave length.

3. Shallow Water. Water that is less in depth than one-half the wave length.

4. Fetch (F). An area of the sea surface over which a wind with a constant direction and speed is blowing, and generating sea waves. The fetch length is measured in nautical miles and has definite boundaries.

5. Duration Time (t). The duration time is the time during which the wind has been in contact with the waves within a fetch.

6. Fully Developed State of the Sea. The fully developed state of the sea is the state the sea reaches when the wind has imparted the maximum energy to the waves.

7. Non-Fully Developed State of the Sea. The non-fully developed state of the sea is the state of the sea reached when the fetch or duration time has limited the amount of energy imparted to the waves by the wind.

8. Steady State. The steady state of the sea is reached when the fetch length has limited the growth of the waves. Once a steady state has been reached, the frequency range produced will not change regardless of the wind.

9. Wind Field. The wind field is a term which refers to the fetch dimensions, wind duration, and wind speed, collectively.

10. Effective Duration Time. The effective duration time is the duration time which has been modified to account for the waves already present in the fetch or to account for waves generated by a rapidly changing wind.

11. E Value. E is equal to the sum of the squares of the individual amplitudes of the individual sine waves which go to make up the actual waves. Since it is proportional to the total energy accumulated in these waves, it is used to describe the energy present in them and in several formulas involving wave energy.

12. Co-cumulative Spectra. The co-cumulative spectra are graphs in which the total accumulated energy is plotted against frequency for a given wind speed. The co-cumulative spectra have been devised for two situations: a fetch limited wind and a duration time limited wind.

13. Upper Limit of Frequencies (f_u). The upper limit of frequencies represents the lowest valued frequencies produced by a fetch or that are present at a forecast point. This term gets its name from the fact that the period associated with this frequency is the period with the highest value. The waves associated with this frequency are the largest waves.

14. Lower Limit of Frequencies (f_L). The lower limit of frequencies represents the highest valued frequencies produced by a fetch or that are present at a forecast point. This term gets its name from the fact that the period associated with this frequency is the period with the lowest value. The waves associated with this frequency are the smallest waves.

15. Filter Area. The filter area is that area between the fetch and the forecast point through which swell waves propagate. This area is so termed because it filters the frequencies and permits only certain ones to arrive at a forecast point at a forecast time.

16. Significant Frequency Range. The significant frequency range is the range of frequencies between the upper limit of frequencies and the lower limit of frequencies. The term significant range is used because those low-valued frequencies whose E values are less than 5 percent of the total E value and those high-valued frequencies whose E values are less than 3 percent of the total E value are eliminated because of insignificance. The significant range of frequencies is

used to determine the range of periods present at the forecast point.

17. Propagation. Propagation, as applied to ocean waves, refers to the movement of the swell through the area between the fetch and the forecast point.

18. Dispersion. Dispersion is the spreading out effect caused by the different group speeds of the spectral frequencies in the original disturbance at the source. Dispersion can be understood by thinking of the different speeds of the different frequencies. The faster wave groups will get ahead of the slower ones; the total area covered is extended thereby. The effect applies to swell only.

19. Angular Spreading. Angular spreading results from waves traveling radially outward from the generating area rather than in straight lines or banks because of different wind directions in the fetch. Although all waves are subject to angular spreading, the effect of such spreading is compensated for only with swell waves because the spreading effect is negligible for sea waves still in the generating area. Angular spreading dissipates energy.

Wave Spectrum

Ocean waves are composed of a multitude of sine waves, each having a different frequency. For explanatory purposes these frequencies are arranged in ascending order from left to right, ranging from the low-valued frequencies on the left to the high-valued frequencies on the right, as illustrated in figure 14-3.

A particular range of frequencies, for instance, from 0.05 to 0.10, does not, however, represent only six different frequencies of sine waves, but rather an infinite number of sine waves whose frequencies range between 0.05 and 0.10. Each sine wave contains a certain amount of energy, and the energy of all the sine waves added together is equal to the total energy present in the ocean waves. The total energy present in the ocean waves is not distributed equally throughout the range of frequencies; instead, in every spectrum, the energy is concentrated around a particular frequency (f_{max}), which corresponds to a certain wind speed. For instance, for a wind speed of 10 knots (kt) f_{max} is 0.248; for 20 kt, 0.124; for 30 kt, 0.0825; for 40 kt, 0.0619. Table 2.1, page 36 in H. O. 603, gives the complete table of f_{max} values and the corresponding periods for wind speeds, starting from 10 kt, at 2-kt intervals. Notice that the frequency decreases as the wind speed increases. This suggests that the higher wind speeds produce higher ocean waves. The table 2.1, mentioned above, can be graphed for each wind speed. An example of such a graph can be found on page 33 of H. O. 603.

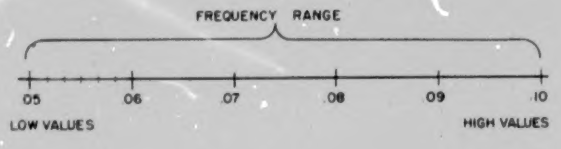

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Figure 14-3.—A typical frequency range of a wave spectrum.

It is difficult to work with actual energy values of these sine waves; for this reason another value has been substituted for energy. This value is the square of the wave amplitude, and it is proportional to wave energy.

The square of the wave amplitude plotted against frequency, for a single value of wind speed, constitutes the spectrum of waves. Thus, a graph of the spectrum is needed for each wind speed, and the energy associated with each sine wave can be determined from these graphs. Each wind speed produces a particular spectrum, and the higher the wind speed, the larger the spectrum.

FORECASTING SEA WAVES

Since sea waves are in the generating area, forecasting of them will generally be most important when units are operating or transiting an area in close proximity to storm centers.

Problems encountered in providing these forecasts will include accurately predicting the storm track and the intensity of the winds that will develop the sea waves.

In this section we will discuss the generation of sea waves, how to most accurately determine wind speeds, and objective methods of forecasting sea waves.

GENERATION AND GROWTH

When the wind starts to blow over a relatively calm stretch of water, the sea surface becomes covered with tiny ripples. These ripples increase in height and decrease in frequency value as long as the wind continues to blow or until a maximum of energy has been imparted to the water for that particular wind speed. These tiny waves are being formed over the entire length and breadth of the fetch. The waves formed near the windward edge of the fetch move through the entire fetch and continue to grow in height and period, so that the waves formed at the leeward edge of the fetch are superimposed on the waves which have come from the windward edge and middle of the fetch. This description illustrates that at the windward edge of the fetch the wave spectrum is small; at the leeward edge of the fetch the spectrum is large.

These waves are generated and grow because of the energy transfer from the wind to the wave. The energy is transferred to the waves by the pushing and dragging forces of the wind. Since the speed of the generated waves is continually increasing, these waves will eventually be traveling at nearly the speed of the wind. When this happens the energy transfer from the wind to the wave ceases. When waves begin to travel faster than the wind, they meet with resistance and lose energy because they are then doing work against the wind. This then explains the limitation of wave height and frequency which a particular wind speed may create.

Fully Developed Sea

When the wind has imparted its maximum energy to the waves, the sea is said to be fully developed. The maximum frequency range for that wind will have been produced by the fetch, and this maximum frequency range will be present at the leeward edge of the fetch. Once the sea is fully developed, no frequency is produced with a value lower than that of the minimum frequency value for the wind speed in question, no matter how long the wind blows. In brief, the waves cannot grow any higher than the maximum value for that wind speed.

When the sea is fully developed, the area near the windward edge is said to be in a steady state, because the frequency range does not increase any more. If the wind continues to blow at the same speed and from the same direction for a considerable period of time, the major portion of the fetch reaches the steady state.

Nonfully Developed Sea

When the wind is unable to impart its maximum energy to the waves, the sea is said to be nonfully developed. This can happen under two circumstances: (1) When the distance over which the wind is blowing is limited; that is, when the fetch is limited; or (2) when the wind has not been in contact with the sea for a sufficient length of time; that is, when the duration time is limited.

FETCH LIMITED SEA.–When the fetch length is too short, the wind is not in contact with the waves over a distance sufficient to impart the maximum energy to the waves. The ranges of frequencies and wave heights are therefore limited and the wave heights are less than those of a fully developed sea. The process of wave generation is cut off before the maximum energy has been imparted to the waves and the fetch is in a steady state. This leads to the conclusion that for every wind speed, a minimum fetch distance is required for the waves to become full developed, and that if this minimum fetch requirement is not met, the sea is fetch limited.

DURATION TIME LIMITED SEA.–When the wind has been in contact with the waves for too short a time, it has had insufficient time to impart the maximum energy to the waves, and the growth of the frequency range and wave heights ceases before the fully developed state of the sea has commenced. Such a situation is known as a duration time limited sea. This leads to the conclusion that for every wind speed, a minimum duration time is required for the waves to become fully developed; and that if this minimum duration time requirement is not met, the sea is duration time limited. The state of the sea, then, is one of three: fully developed, fetch limited, or duration time limited.

Table 14-1 shows the minimum wind duration times and fetch lengths needed to generate a fully developed state of the sea for various wind speeds. When the actual conditions do not meet these minimum requirements, the properties of the waves must be determined by means of graphs and formulas.

DETERMINING THE WIND FIELD

As we have discussed, wind is the cause of waves. It therefore stands to reason that in order to accurately predict sea conditions it is necessary to determine wind properties as accurately as possible. Miscalculation of fetch, wind speed, or duration will only lead to inaccuracies in predicted wave conditions.

In this section methods of determining the wind properties, as accurately as possible with data available, an presented.

Location of Fetch

In all cases the first step toward a wave forecast is locating a fetch. A fetch is an area of the sea surface over which a wind with a constant direction and speed is blowing. Figure 14-4 shows some typical fetch areas. The ideal fetch over the open ocean is rectangular, with the winds constant in both speed and direction. As shown in figure 14-4, most fetch areas are bounded by coastlines, frontal zones or a change in isobars. In cases where the curvature of the isobars is large, it is a good practice to use more than one fetch area, as shown in figure 14-4(B).

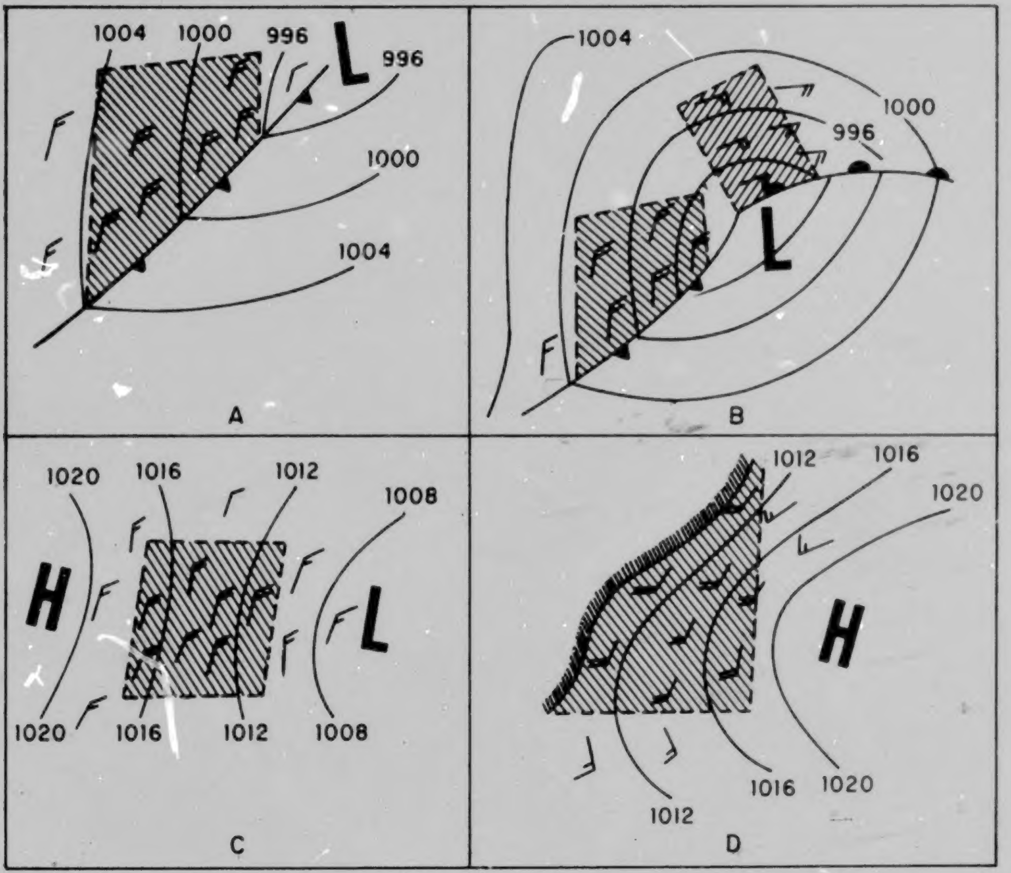
Although some semipermanent pressure systems have stationary fetch areas, and some storms may move in such a manner that the fetch is practically stationary, there are also many moving fetch areas. Figure 14-5 shows 3 cases where a fetch AB has moved to the position CD on the next map 6 hours later. The problem is determining what part of the moving fetch area to consider as the average fetch for the 6 hour period.

In figure 14-5, Case 1, the fetch moves perpendicular to the wind field. The best approximation is fetch CB. Therefore in a forecast involving this type of fetch, use only the part of the fetch that appears on two consecutive maps. The remaining fetch does contain waves, but they are lower than those in the overlap area.

In figure 14-5, Case 2, the fetch moves to leeward (in the same direction as the wind). Since waves are moving forward through the

Table 14-1.—Minimum fetch length (in nautical miles) and minimum duration time (in hours needed to generate a fully developed sea versus wind speed (in knots).

V	F	t
10	10	2.4
12	18	3.8
14	28	5.2
16	40	6.6
18	55	8.3
20	75	10
22	100	12
24	130	14
26	180	17
28	230	20
30	280	23
32	340	27
34	420	30
36	500	34
38	600	38
40	710	42
42	830	47
44	960	52
46	1,100	57
48	1,250	63
50	1,420	69
52	1,610	75
54	1,800	81
56	2,100	88


1004
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H
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1020
1012
D

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Figure 14-4.—Typical fetch areas.

fetch area, the area to be used in this case is fetch CD.

Case 3, figure 14-5, depicts the fetch moving windward (against the wind). Since the waves move toward A, the region AC will have higher waves than the area BD. Experience has shown that in this case AB is the most accurate choice for a fetch.

Determining Accurate Wind Speed

The most obvious and accurate way to determine wind speed over a fetch is to average the reported values from ships. This method has the advantage of not requiring a correction for gradients or stability. However, realistically, more often there are only a few ship reports available and ship reports are subject to error in observation, encoding, or transmission.

A second way to determine wind speed is to measure the geostrophic wind from the isobaric spacing and then correct it for curvature and stability. At first it would seem this would be less desirable due to the extra time making corrections. However, barometric pressure is probably the most reliable of the parameters

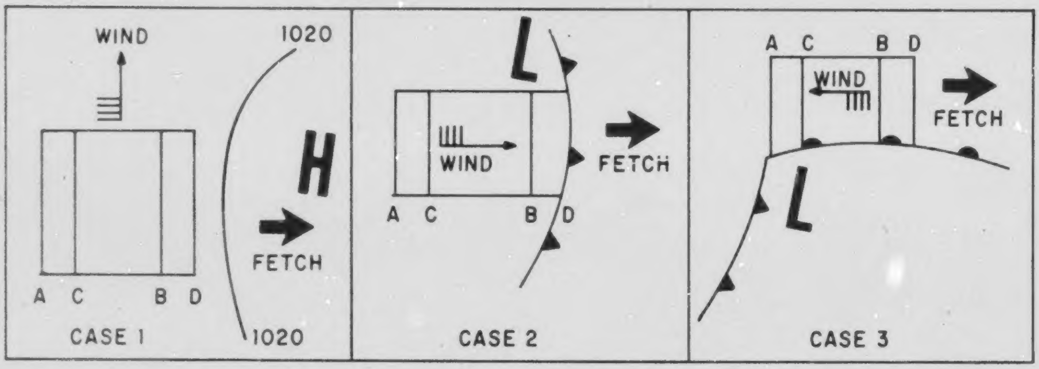

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Figure 14-5.—Examples of moving fetches.

reported by ships and a reasonably accurate isobaric analysis can be made from a minimum number of reports. For these reasons the corrected geostrophic wind is considered to be the best measure of wind speed over the fetch, except of course in cases where there is a dense network of ship reports where wind direction and speed are in good agreement.

The reason for correction to geostrophic wind is that the isobars must be straight for a correct measure of the wind. When the isobars curve, other forces enter into the computations. The wind increases or decreases depending on whether the system is cyclonic or anticyclonic in nature. The stability correction is a measure of the turbulence in the layer above the water. Cold air over warmer water is unstable and highly turbulent, making the surface wind more nearly equal to the geostrophic wind. Conversely, warm air over colder water produces a stable air mass and results in the surface wind being much smaller than the geostrophic wind.

Three rules for an approximation of the curvature correction are as follows:

1. For moderately curved to straight isobars—no correction is applied.
2. For great anticyclonic curvature—add 10 percent to the geostrophic wind speed.
3. For great cyclonic curvature—subtract 10 percent from the geostrophic wind speed.

In the majority of cases the curvature correction can be neglected since isobars over a fetch area are relatively straight. The gradient wind can always be computed if more refined computations are desired.

In order to correct for air mass stability the sea-air temperature difference must be computed. This can be done from ship reports in or near the fetch area aided by climatic charts of average monthly sea surface temperatures when data is too scarce. The correction to be applied is given in table 14-2. The symbol Ts stands for the temperature of the sea surface, and Ta for the air temperature.

Table 14-2.—Air-sea temperature difference correction.

(T_s - T_a) Algebraically Subtracted	Percent of Geostrophic Wind
0 or negative	60
0 to 10	65
10 to 20	75
20 or above	90

Determination of Wind Duration

Once a fetch has been determined and the wind speed has been found, the next step is to determine the duration of that wind over the fetch. It is highly unlikely that the wind will begin and end at one of the 6 hour map times. Therefore an accurate value must be interpolated. In most cases a simple interpolation of the successive maps will be sufficient to locate the bounds of the wind field in space and time.

Determining how long the wind has blown is relatively simple when the wind speed has been constant for the entire duration. If this does not occur, a representative duration must be selected.

SLOWLY VARYING WIND. Suppose the wind has been blowing for 24 hours, with velocities of 10 knots for 6 hours, 15 knots for 12 hours, and 20 knots for 6 hours. The duration is 24 hours but the speed value is in question. The most consistent solution is to use three durations with the corresponding wind speeds and work up three successive states.

MORE RAPID VARIATIONS. Suppose the wind blows for 12 hours and during that time increases in velocity from 10 to 20 knots. Studies and experience have shown that in cases of variable winds a single value may be assigned for wind speed if the change has been relatively small. The following rules can be applied under these conditions:

1. Average the wind speeds when the change is gradual or increasing, then decreasing. Apply the average to the entire duration.
2. Use the last wind speed when the speed changes in the first few hours, then remains constant. Apply that speed to the entire duration.

OBJECTIVE METHODS FOR FORECASTING SEA WAVES

There are a number of different methods for forecasting sea waves. Some of the methods are too technical or time consuming to be of practical use to Aerographer's Mates.

One of the most complete methods is the Pierson-Neuman-James Method in H. O. 603. The Oceanographic Services Section, Naval Weather Service Facility, San Diego, has devised a reduced version of this method, incorporating the use of graphs and worksheets. Personnel with limited knowledge of technical theory can become proficient in its use with experience. This method will be discussed in this manual.

Figure 14-6 illustrates the worksheet which will be used with this forecasting technique. The step-by-step instructions which follow coincide with those on the worksheet.

STEP 1. Determine the average wind speed (U) over the fetch and enter this value on the worksheet. Keep in mind that the basic definition of a fetch requires that the wind be the same speed throughout and the value of wind speed should be very close to the actual wind at any point in the fetch.

STEP 2. Measure the length of the fetch in nautical miles and enter on the worksheet as (F).

STEP 3. Determine how long the wind has maintained the same speed as (U) by checking back through previous weather charts. Enter the number of hours as (t_d) on the worksheet.

STEP 4. Turn to sea and swell graphs 1a and 1b (figures 14-7 and 14-8). These are two C. C. S. duration graphs. Graph 1a is for wind speeds of 10 to 44 knots and 1b is for wind speeds of 36 to 56 knots. Use appropriate graph for the wind speed.

Enter the graph with the wind speed (U) from Step 1 and the duration of the wind (t_d) from Step 3. At the intersection of the wind speed and duration, go horizontally to the left scale and read the energy at intersection (E_i). Record this value on the worksheet. From the same intersection go to the bottom scale and read the frequency at intersection (f_i). Record this value on the worksheet.

STEP 5. Turn to sea and swell graphs 1c and 1d (figures 14-9 and 14-10). These are two C. C. S. fetch graphs. Graph 1c is for wind speeds of 10 to 44 knots and 1d is for winds of 36 to 56 knots. Use the appropriate graph for wind speed (U) found in Step 1.

Enter the graph with the wind speed (U) and fetch length (F). At the intersection of these values go horizontally to the left scale and read the energy at intersection (E_i). Record this value on the worksheet. From the same intersection go to the bottom scale and read the frequency at intersection (f_i). Record this value on the worksheet.

SEA WORKSHEET

FETCH # ________ DTG CHART ______/______________ DATE ____________

* * *

Observed or Forecast Parameters

1. Wind Speed over Fetch $U =$ __________ kts.

2. Fetch Length $F =$ __________ mi.

3. Duration of Wind $t_d =$ __________ hrs.

* * *

Graphical Calculations

Step No.	Enter Sea & Swell Graph	With	And Read	
4*	1a or 1b	U from step 1 and t_d from step 3.	E_i f_i	ft^2 cps
5*	1c or 1d	U from step 1 and F from step 2.	E_i f_i	ft^2 cps
6	None	Choose the smaller E_i from steps 4 & 5, along with its corresponding f_i.	E_i f_i	ft^2 cps
7	None	If the f_i chosen in step 6 lies to the left of f_{max} on the CCS Graph, then f_u is the same as f_i. If not, go on to next step.	f_u	cps
8	2	If f_i chosen in step 6 lies to the right of f_{max} on the CCS Graph, enter Graph 2 with f_i and read f_u	f_u	cps
9	3	E_i from step 6.	$.03E_i$	ft^2
10	1 a,b,c, or d	U from step 1 and $.03E_i$ from step 9.	f_L	cps
11	4	f_u from step 7 or 8. Repeat for f_L from step 10.	T_u T_L	sec sec
12	5	U from step 1 and f_i from step 6.	T_{avg}	sec
13	6a or 6b	E_i from step 6.	H_{avg} $H_{1/3}$ $H_{1/10}$	ft ft ft

*If no intersection of U and t_d or U and F occurs with the values found in steps 1, 2, and 3, then the sea is fully developed. Go to the fully developed sea table and read T_u, T_L, T_{avg}, H_{avg}, H_{v3}, and $H_{1/10}$ directly.

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Figure 14-6.—Sea worksheet.

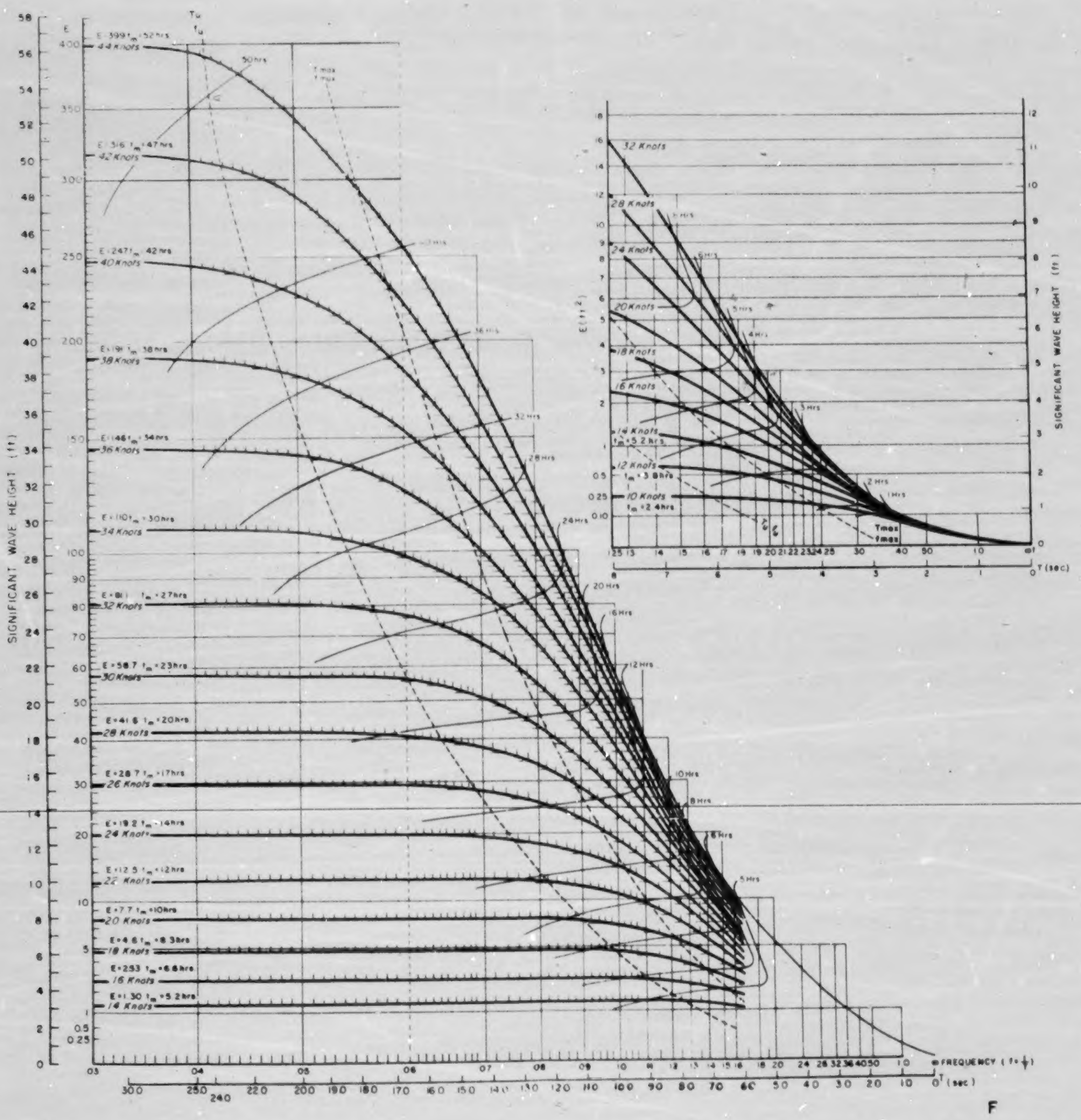

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Figure 14-7.—Sea and swell graph 1a. Distorted C. C. S. (duration graph-wind speeds 10-44 knots).

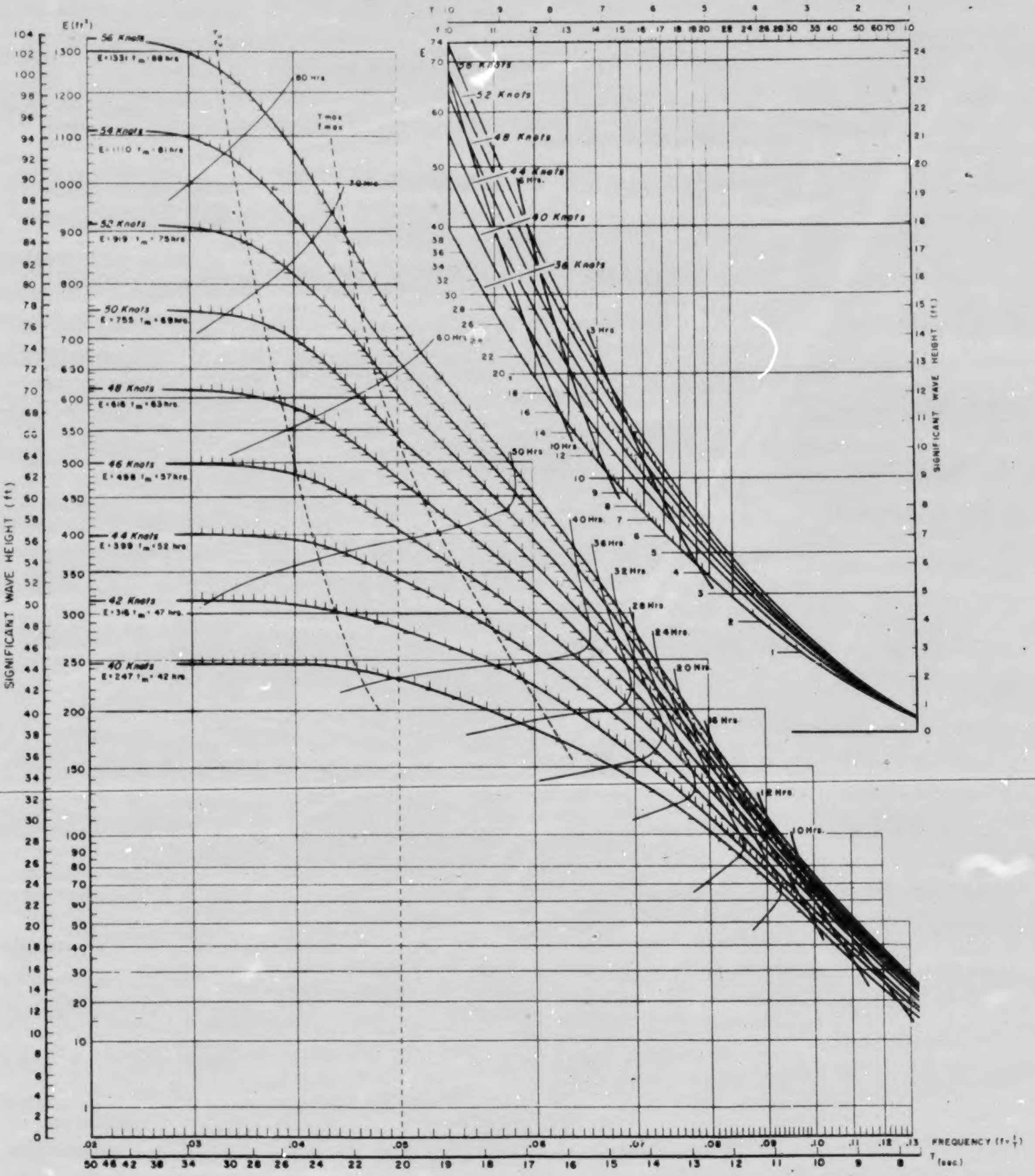

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Figure 14-8.—Sea and swell graph 1b. Distorted C. C. S. (duration graph-wind speeds 36-56 knots).

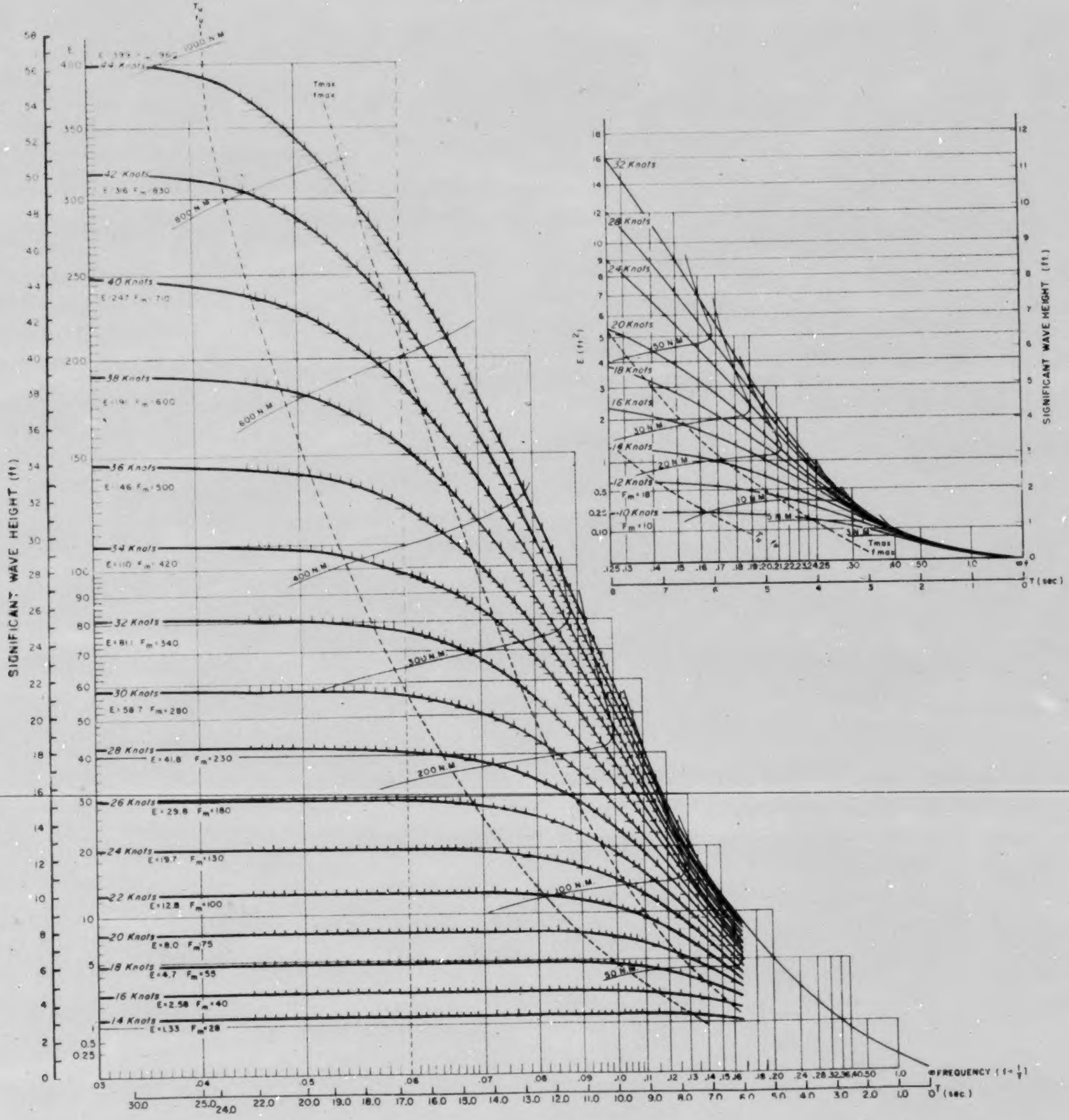

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Figure 14-9.—Sea and swell graph 1c. Distorted C. C. S. (fetch graph-wind speeds 10-44 knots).

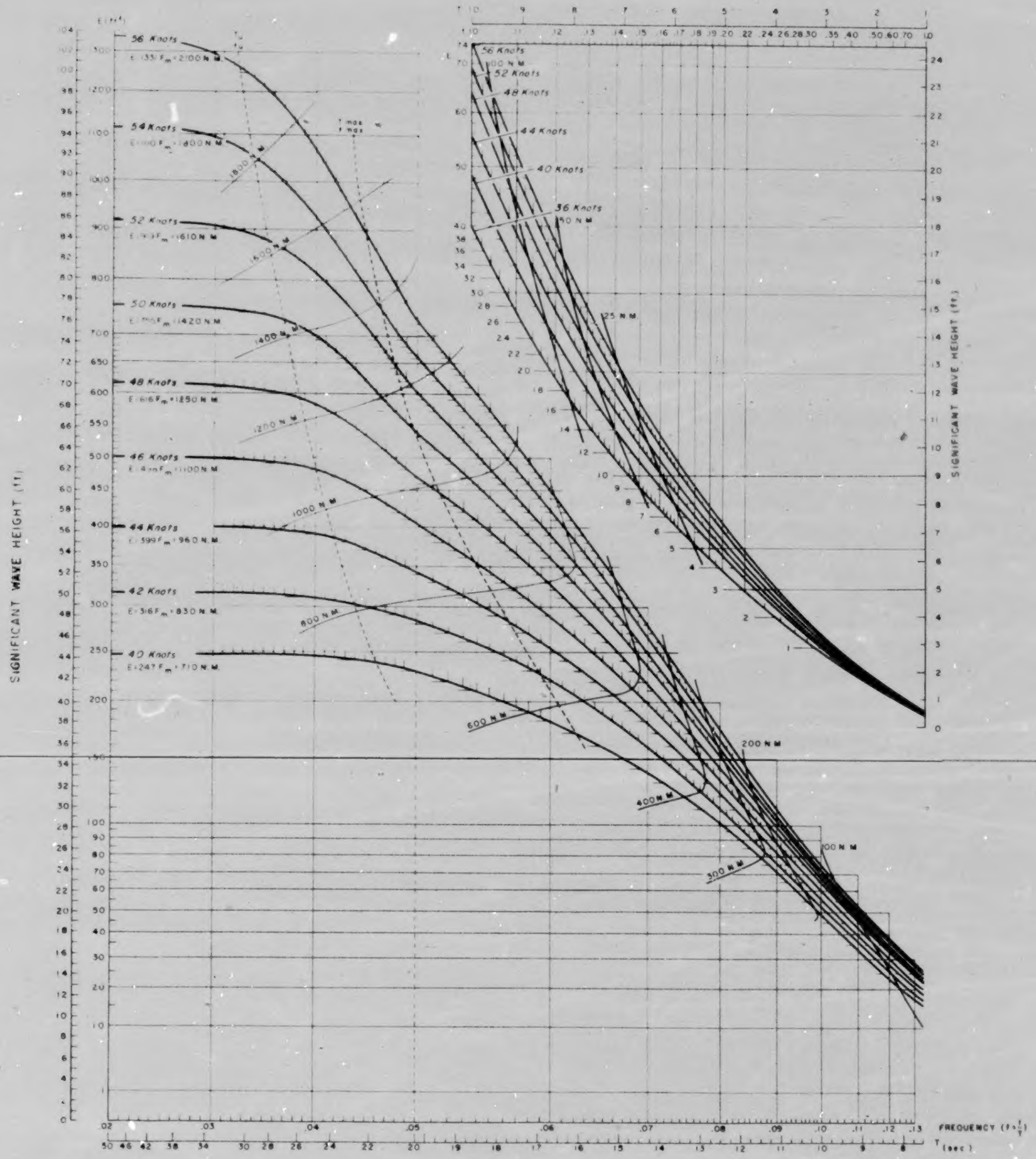

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Figure 14-10.—Sea and swell graph 1d. Distorted C. C. S. (fetch graph-wind speeds 36-56 knots).

STEP 6. Examine the E_i values found in Step 4 and Step 5. Chose the smaller value of E_i and its associated f_i. Record these values for Step 6 on the worksheet, and use these values as required for all further steps.

NOTE: The duration line on the C. C. S. duration graph may not be long enough to intersect the wind speed line. DO NOT EXTEND OR EXTRAPOLATE THE DURATION LINE IN ANY WAY. The fact that the duration line does not reach all the way to the wind speed line means that the sea state is not duration limited. In this case there is no intersection and no value of E_i or f_i can be recorded on the worksheet.

The same situation can occur with the fetch length line on the fetch graphs. If so, it means that the sea state is not fetch limited.

If only one of the graphs has an intersection use that value of E_i and f_i for further steps without concern for a smaller E_i.

If neither of the graphs has an intersection, the sea is fully developed and no further computations are required. Go directly to table 14-3 for fully developed sea and read the values for period (T) and wave height (H).

STEP 7. If the intersection chosen in Step 6 lies to the left of the dashed line labeled f_{max} then the upper frequency (f) will be the same as the value of f_i found in Step 6. Record the value of f_i as f_u on the worksheet. If the intersection lies to the right of the f_{max} line, then go on to Step 8 to determine f_u.

STEP 8. If the intersection chosen in Step 6 lies to the right of the line labeled f_{max}, then turn to sea and swell graph 2, figure 14-11.

On the skeletonized C. C. S. graph depicted in figure 14-12, the intersection lies to the right of the f_{max} line.

Enter the left side of sea and swell graph 2 with the frequency at intersection (f_i) and go across horizontally to the diagonal line. From this point move vertically downward to the bottom of the graph and read the value of the upper frequency (f_u). Record this value on the worksheet.

STEP 9. Turn to sea and swell graph 3, figure 14-13. Enter the left side of the graph with the energy at intersection (E_i) found in Step 6. Go horizontally to the right to the bent diagonal line. From this point move vertically downward to the bottom scale of the graph and read the value of 3 percent of E. Record this value in the space marked $.03E_i$ on the worksheet.

STEP 10. Return to the C. C. S. graph having the intersection found in Step 6. (This will be one of the sea and swell graphs, figure 14-7, 14-8, 14-9, or 14-10.) Enter the left side of the graph at the E scale, with the value of $.03E_i$ found in Step 9. Go horizontally to the wind speed line (U) found in Step 1. From that point move vertically downward to the f scale at the bottom of the graph and read the value at the lower frequency (f_L). Record this value on the worksheet.

STEP 11. Turn to sea and swell graph 4, figure 14-14. Enter the left side of the graph with the value of upper frequency (f_u) found in step 8. Go horizontally across to the curved line, then vertically downward to the bottom scale. Read the value of the upper period (T_u) and record it on the worksheet.

STEP 12. Turn to sea and swell graph 5, figure 14-15. Enter the graph with the values of wind speed (U) found in Step 1 and frequency at intersection (f_i) found in Step 6. Instructions for the use of this graph are found in the lower right corner. Follow these directions and read the value of the average period (T_{avg}). Record this value on the worksheet.

STEP 13. Turn to sea and swell graphs 6a and 6b, figures 14-16 and 14-17. These graphs are the same except that 6b is a blow-up version of 6a, to give a more accurate result when working with small values. Enter the graph with the energy at intersection (E_i) found in Step 6. If the E_i value is less than 10 ft^2 use graph 6b (fig. 14-17). If greater than 10 ft^2 use graph 6a (fig. 14-16). Follow the instructions printed at the top of the graph to get characteristic wave heights. This will have to be repeated three times; once to get H_{avg}; once to get H; and once to get H. Record these values on the worksheet.

Table 14-3.–Wind and sea scale for fully arisen sea.

WIND AND SEA SCALE FOR FULLY ARISEN SEA														
SEA-GENERAL		WIND				SEA								
						WAVE HEIGHT FEET								
SEA STATE	DESCRIPTION	BEAUFORT WIND FORCE	DESCRIPTION	RANGE (KNOTS)	WIND VELOCITY (KNOTS)	AVERAGE	SIGNIFICANT	AVERAGE 1/10 HIGHEST	SIGNIFICANT RANGE OF PERIODS (SECONDS)	(PERIOD OF MAXIMUM ENERGY OF SPECTRUM)	AVERAGE PERIOD	AVERAGE WAVE LENGTH	MINIMUM FETCH (NAUTICAL MILES)	MINIMUM DURATION (HOURS)
0	Sea like a mirror.	0	Calm	Less than	0	0	0	0	-	-	-	-	-	-
	Ripples with the appearance of scales are formed, but without foam crests.	1	Light Airs	1 - 3	2	0.05	0.08	0.10	up to 1.2 sec	0.7	0.5	10 in.	5	18 min
1	Small wavelets, still short but more pronounced; crests have a glassy appearance, but do not break.	2	Light Breeze	4 - 6	5	0.18	0.29	0.37	0-4-2.8	2.0	1.4	6.7 ft	8	39 min
	Large wavelets, crests begin to break. Foam of glassy appearance. Perhaps scattered white horses.	3	Gentle Breeze	7 - 10	8.5	0.6	1.0	1.2	0-8-5.0	3.4	2.4	20	9.8	1.7 hrs
					10	0.88	1.4	1.8	1.0-6.0	4	2.9	27	10	2.4
2	Small waves, becoming larger; fairly frequent white horses.	4	Moderate Breeze	11-16	12	1.4	2.2	2.8	1.0-7.0	4.8	3.4	40	18	3.8
					13.5	1.8	2.9	3.7	1.4-7.6	5.4	3.9	52	24	4.8
3					14	2.0	3.3	4.2	1.5-7.8	5.6	4.0	59	28	5.2
					16	2.9	4.6	5.8	2.0-8.8	6.5	4.6	71	40	6.6
4	Moderate waves, taking a more pronounced long form; many white horses are formed. (Chance of some spray).	5	Fresh Breeze	17-21	18	3.8	6.1	7.8	2.5-10.0	7.2	5.1	90	55	8.3
					19	4.3	6.9	8.7	2.8-10.6	7.7	5.4	99	65	9.2
					20	5.0	8.0	10	3.0-11.1	8.1	5.7	111	75	10
5	Large waves begin to form; the white foam crests are more extensive everywhere. (Probably some spray).	6	Strong Breeze	22-27	22	6.4	10	13	3.4-12.2	8.9	6.3	134	100	12
					24	7.9	12	16	3.7-13.5	9.7	6.8	160	130	14
6					24.5	8.2	13	17	3.8-13.6	9.9	7.0	164	140	15
					26	9.6	15	20	4.0-14.5	10.5	74	188	180	17
	Sea heaps up and white foam from breaking waves begins to be blown in streaks along the direction of the wind. (Spindrift begins to be seen).	7	Moderate Gale	28-33	28	11	18	23	4.5-15.5	11.3	7.9	212	230	20
7					30	14	22	28	4.7-16.7	12.1	8.6	250	280	23
					30.5	14	23	29	4.8-17.0	12.4	8.7	258	290	24
					32	16	26	33	5.0-17.5	12.9	9.1	285	340	27
	Moderately high waves of greater length; edges of crests break into spindrift. The foam is blown in well marked streaks along the direction of the wind. Spray effects visibility.	8	Fresh Gale	34-40	34	19	30	38	5.5-18.5	13.6	9.7	322	420	30
					36	21	35	44	5.8-19.7	14.5	10.3	363	500	34
					37	23	37	46.7	6. 20.5	14.9	10.5	376	530	37
					38	25	40	50	6.2-20.8	15.4	10.7	392	600	38
					40	28	45	58	6.5-21.7	16.1	11.4	444	710	42
8	High waves. Dense streaks of along the direction of the wind. Sea begins to roll. Visibility affected.	9	Strong Gale	41-47	42	31	50	64	7-23	17.0	12.0	492	830	47
					44	36	58	73	7-24.2	17.7	12.5	534	960	52
					46	40	64	81	7-25	18.6	13.1	590	1110	57
9	Very high waves with long overhanging crests. The resulting foam is in great patches and is blown in dense white streaks along the direction of the wind. On the whole the surface of the sea takes a white appearance. The rolling of the sea becomes heavy and shock-like. Visibility is affected.	10	Whole Gale	48-55	48	44	71	90	7.5-26	19.4	13.8	650	1250	63
					50	49	78	99	7.5-27	20.2	14.3	700	1420	69
					51.5	52	83	106	8-28.2	20.8	14.7	736	1560	73
					52	54	87	110	8-28.5	21.0	14.8	750	1610	75
					54	59	95	121	8-29.5	21.8	15.4	810	1800	81
	Exceptionally high waves (Small and medium-sized ships might for a long time be lost to view behind the waves.) The sea is completely covered with long white patches of foam lying along the direction of the wind. Everywhere the edges of the wave crests are blown into froth. Visibility affected.	11	Storm	56-63	56	64	103	130	8-5-31	22.6	16.3	910	2100	88
					59.5	73	116	148	10-32	24	17.0	985	2500	101
	Air filled with foam and spray. Sea completely white with driving spray; visibility very seriously affected.	12	Hurricane	64-71	>64	>80	>128	>164	101(35)	(26)	(18)	~	~	~

The objective technique of forecasting sea waves is completed with determining of sea height computations in the preceding manner.

The manner of presentation of the wave information to the user can be determined on the local level to the satisfaction of the organizations involved.

FORECASTING SWELL WAVES

In the preceding portion of this chapter we have discussed the principles of sea waves and methods of forecasting them. With sea wave forecasting we are considering the point for which we are forecasting to be within the generating area, with the wind still blowing. This, however, will not be the problem in the majority of the forecasts that will be required. Normally the point for which the forecast is prepared will be outside of the fetch area; therefore it will be necessary to determine what effect the distance traveled is going to have on the waves. In this section we will discuss the

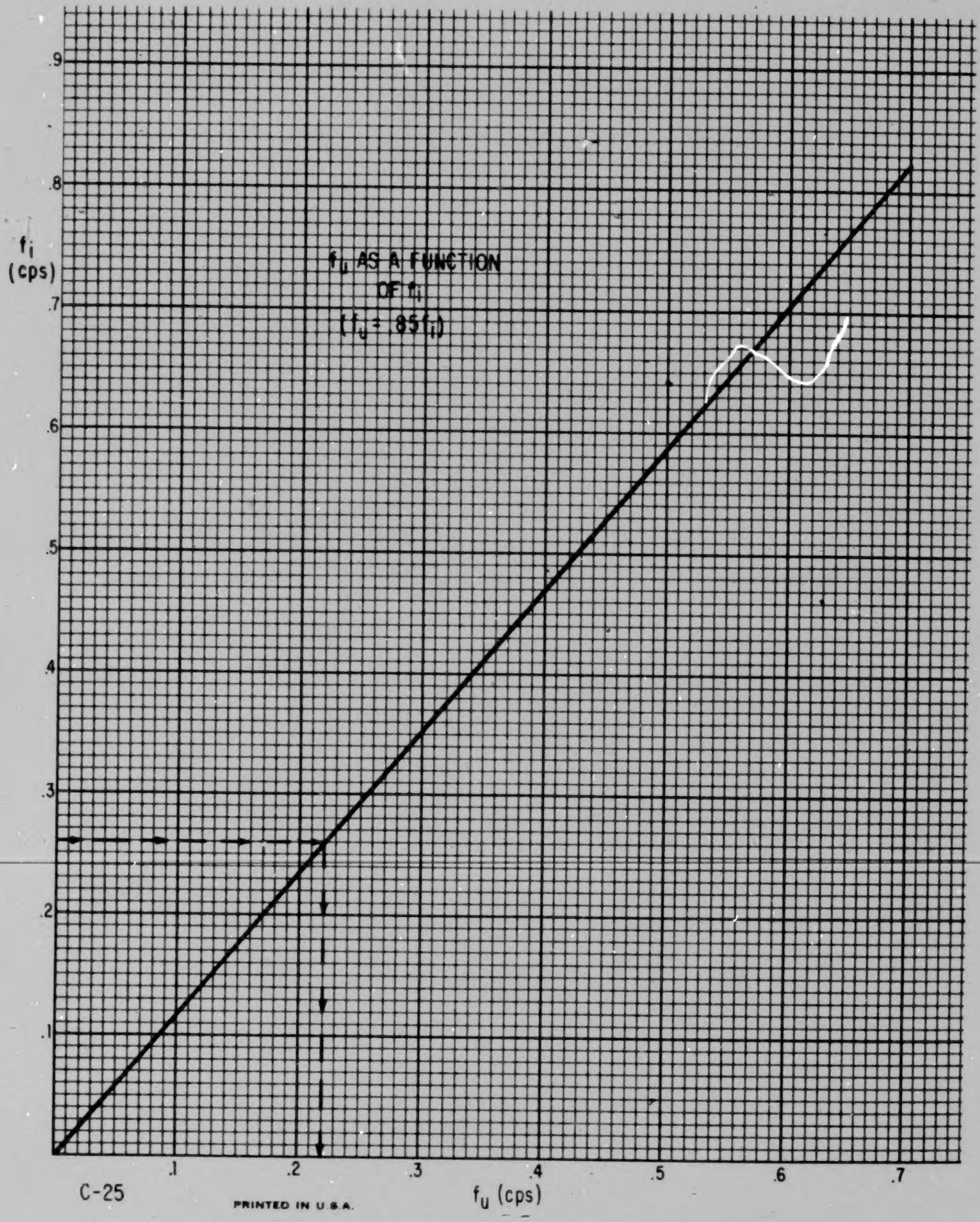

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Figure 14-11.—Sea and swell graph 2.

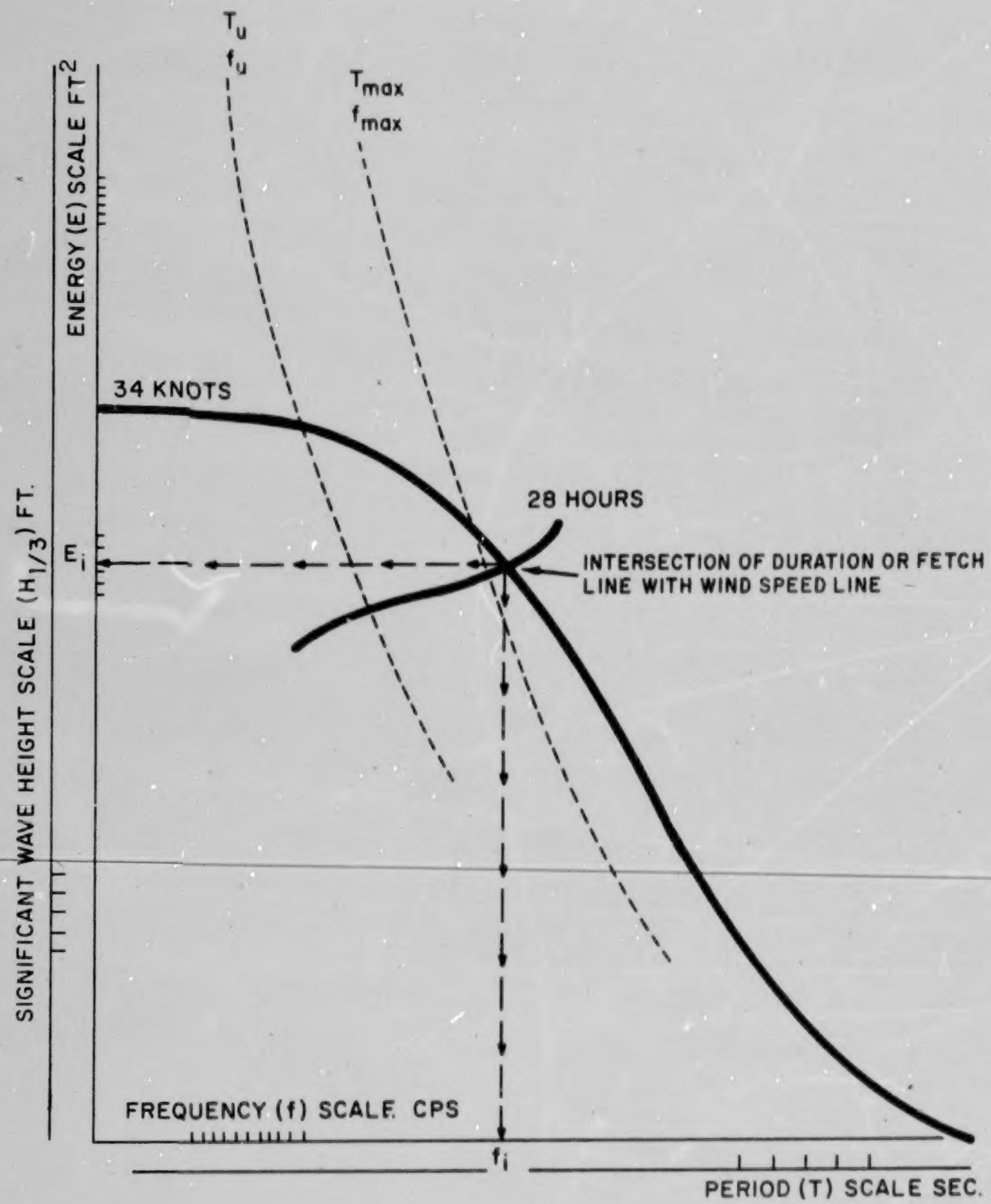

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Figure 14-12.–Skeletonized C. C. S. graph showing major lines.

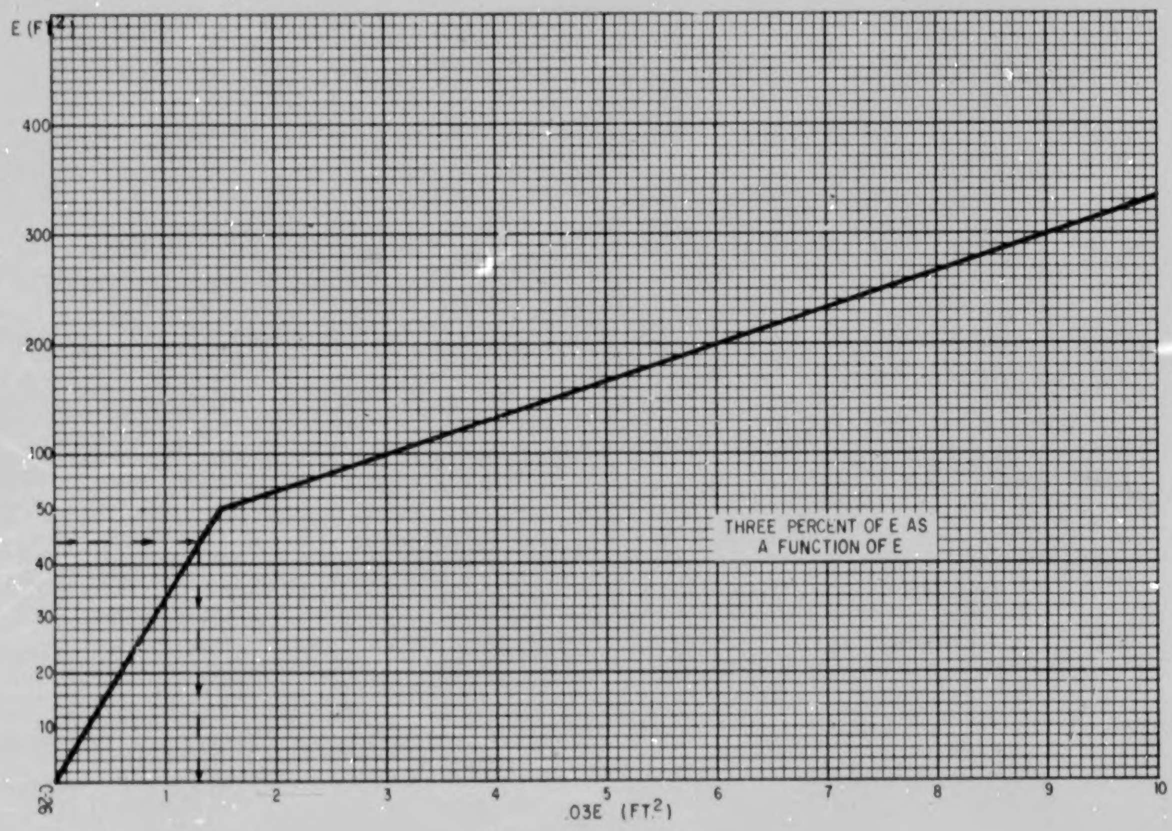

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Figure 14-13.—Sea and swell graph 3.

basic principles of swell waves as well as an objective method of determining what changes will take place in the spectrum of waves as they traverse from the generating area to the forecast point.

GENERATION OF SWELL WAVES

After a sea state has been generated in a fetch, there are many different wave trains present with different periods, most of which are moving out of the fetch in slightly different directions. Because of these different periods and slight differences in direction, the propagation of swell waves follows two fundamental processes. These processes are dispersion and angular spreading.

Dispersion

An accepted fact about wave travel is that the waves with higher periods move faster than waves with shorter periods. The actual formula for the speed of the wave train is

$$C = 1.515T$$

where C is the speed of the wave train and T is the wave period in the wave train.

All of the different wave trains (series of waves all having the same period and direction of movement) in the fetch can be compared to a group of long distance runners at a track and field meet. At first all of the runners start out at the starting line at the same time. As they

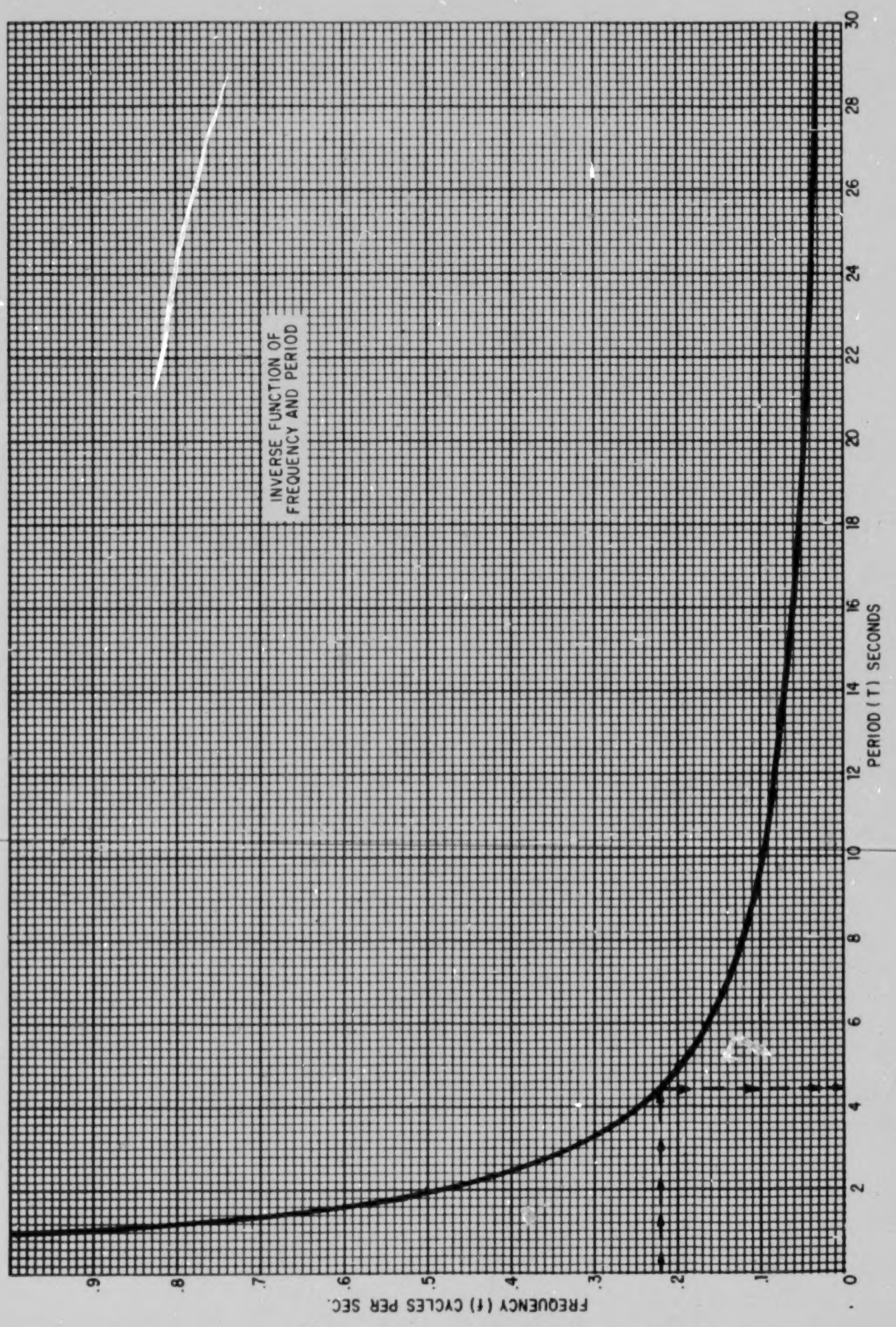

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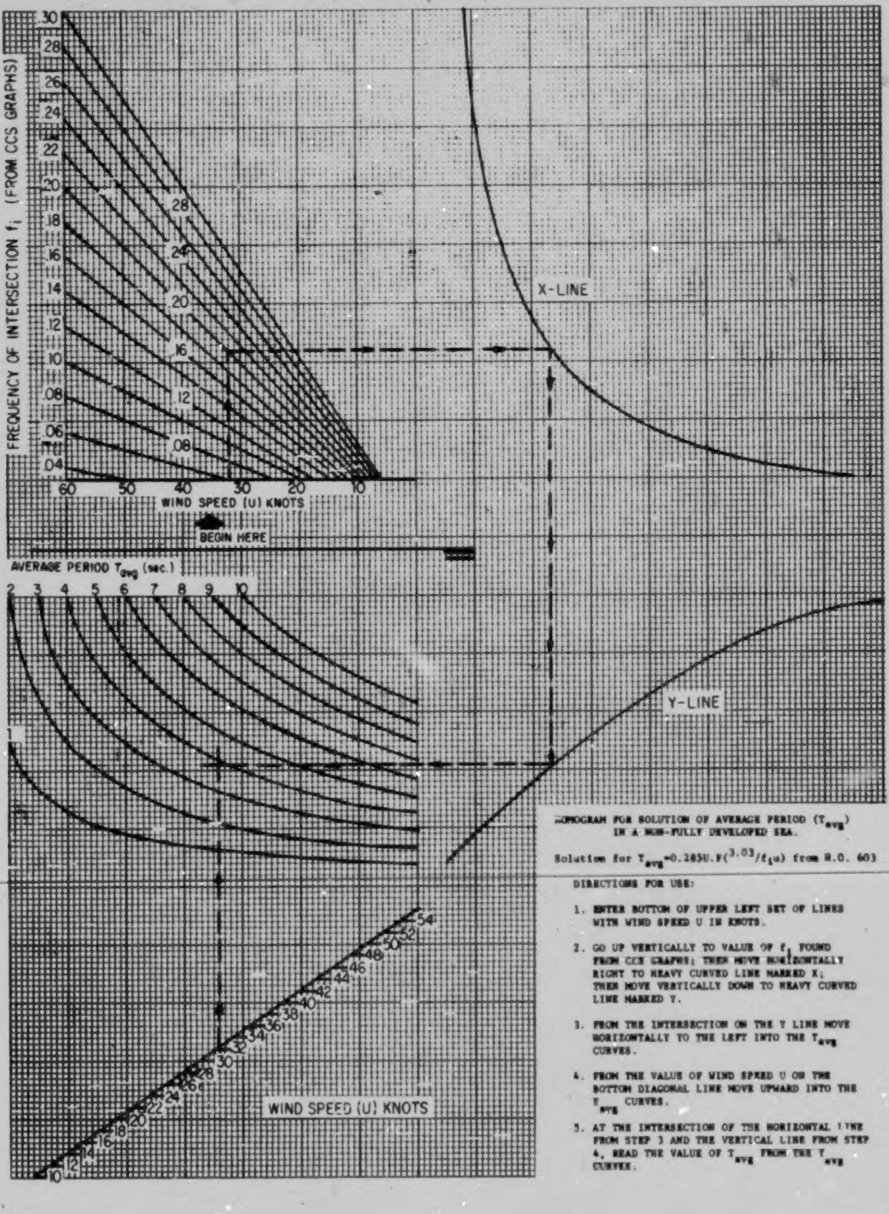
Figure 14-14.—Sea and swell graph 4.



NOMOGRAM FOR SOLUTION OF AVERAGE PERIOD (T_{avg})
IN A NON-FULLY DEVELOPED SEA.

Solution for $T_{avg}=0.285U.F(^{3.03}/f_{i}u)$ from H.O. 603

DIRECTIONS FOR USE:

1. ENTER BOTTOM OF UPPER LEFT SET OF LINES WITH WIND SPEED U IN KNOTS.
2. GO UP VERTICALLY TO VALUE OF f_i FOUND FROM CCS GRAPHS; THEN MOVE HORIZONTALLY RIGHT TO HEAVY CURVED LINE MARKED X; THEN MOVE VERTICALLY DOWN TO HEAVY CURVED LINE MARKED Y.
3. FROM THE INTERSECTION ON THE Y LINE MOVE HORIZONTALLY TO THE LEFT INTO THE T_{avg} CURVES.
4. FROM THE VALUE OF WIND SPEED U ON THE BOTTOM DIAGONAL LINE MOVE UPWARD INTO THE T_{avg} CURVES.
5. AT THE INTERSECTION OF THE HORIZONTAL LINE FROM STEP 3 AND THE VERTICAL LINE FROM STEP 4, READ THE VALUE OF T_{avg} FROM THE T_{avg} CURVES.

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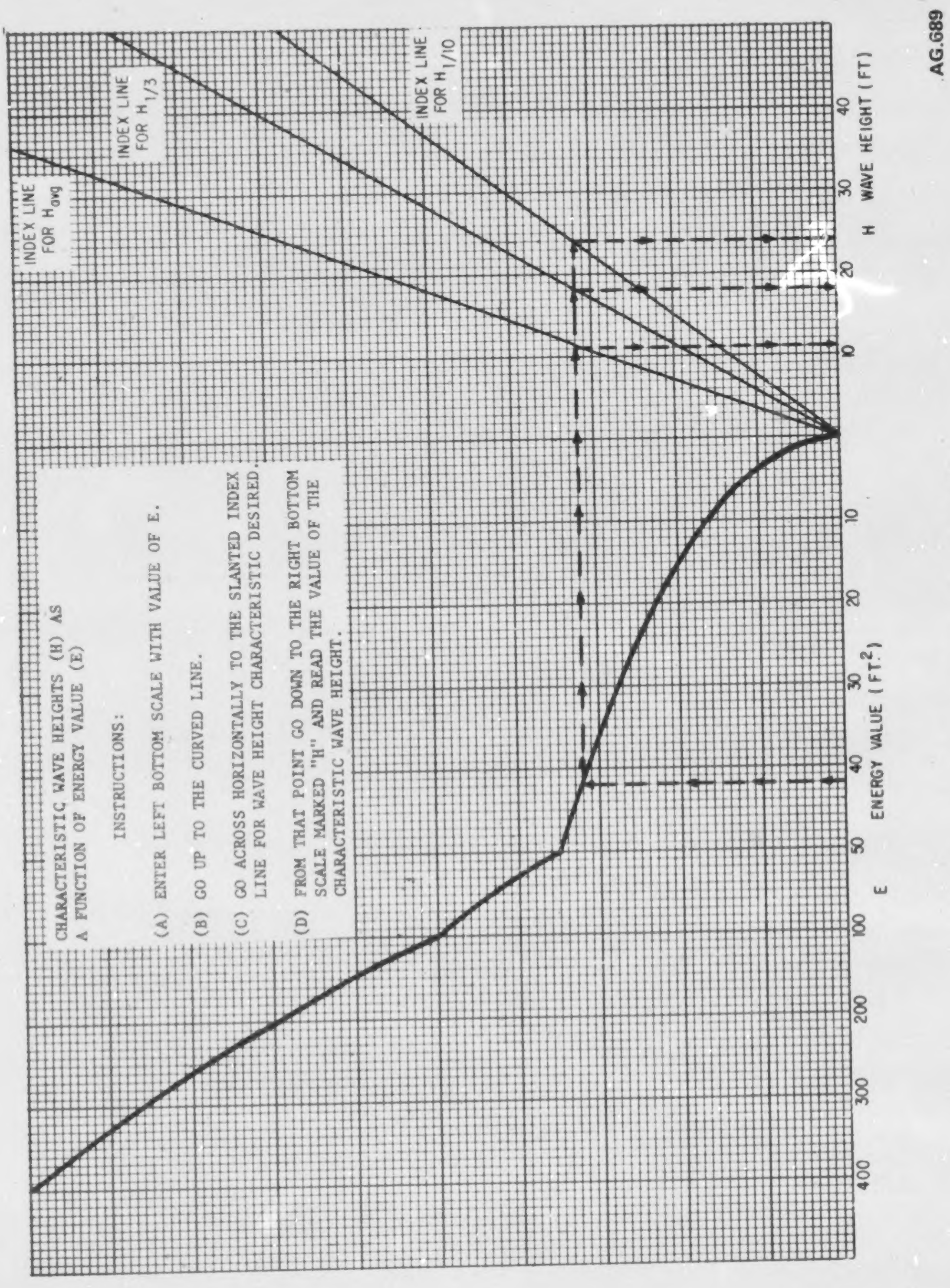
Figure 14-15.—Sea and swell graph 5.



CHARACTERISTIC WAVE HEIGHTS (H) AS
A FUNCTION OF ENERGY VALUE (E)

INSTRUCTIONS:

(A) ENTER LEFT BOTTOM SCALE WITH VALUE OF E.

(B) GO UP TO THE CURVED LINE.

(C) GO ACROSS HORIZONTALLY TO THE SLANTED INDEX LINE FOR WAVE HEIGHT CHARACTERISTIC DESIRED.

(D) FROM THAT POINT GO DOWN TO THE RIGHT BOTTOM SCALE MARKED "H" AND READ THE VALUE OF THE CHARACTERISTIC WAVE HEIGHT.

Figure 14-16.—Sea and swell graph 6a.

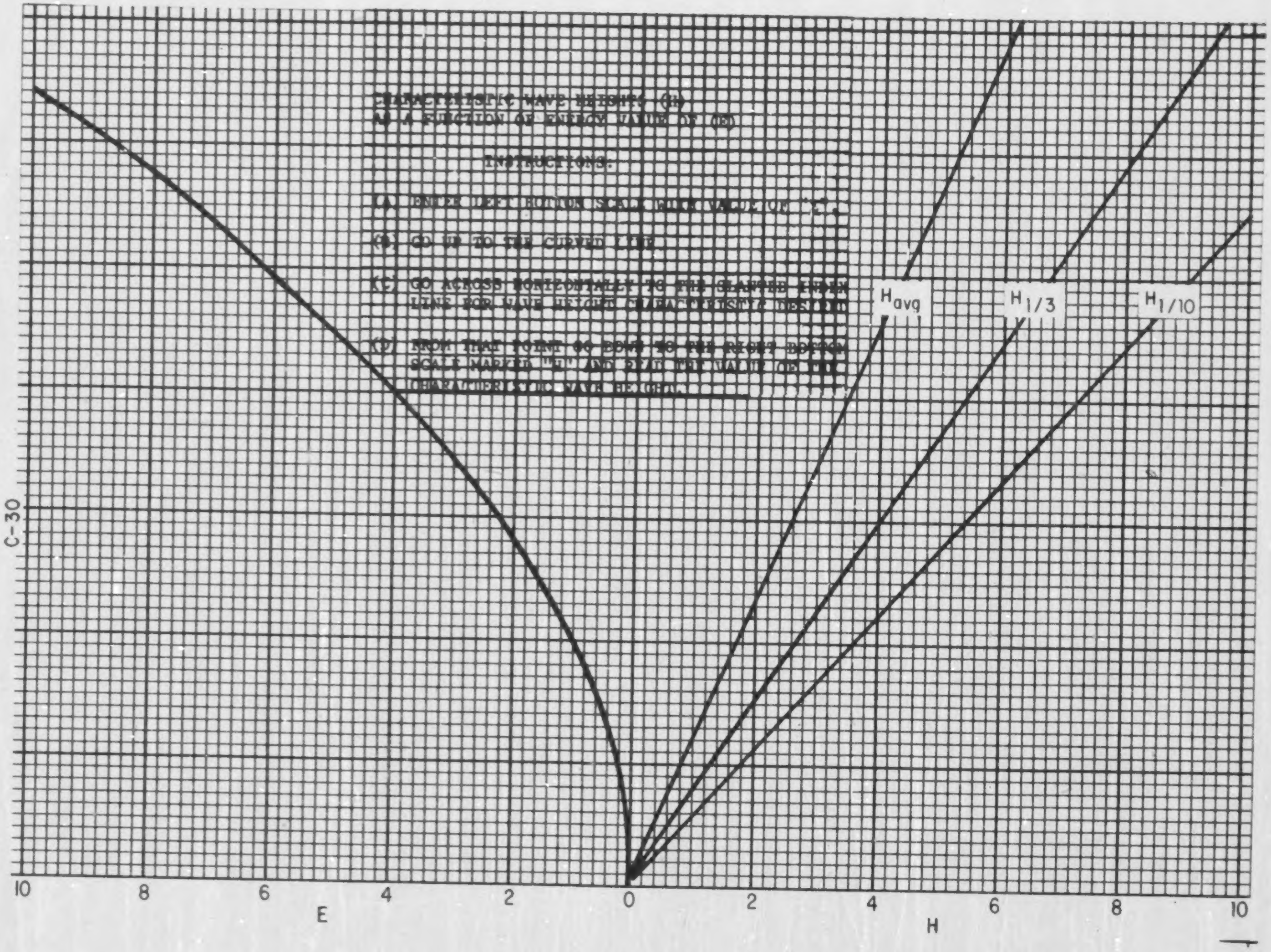


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Figure 14-17.—Sea and swell graph 6b.

Frequency	Period	Distance traveled in nautical miles	Length of disturbance in nautical miles
.050	20	1960	300
.067	15	1470	225
.083	12	1180	180
.100	10	980	150
.167	6	590	90

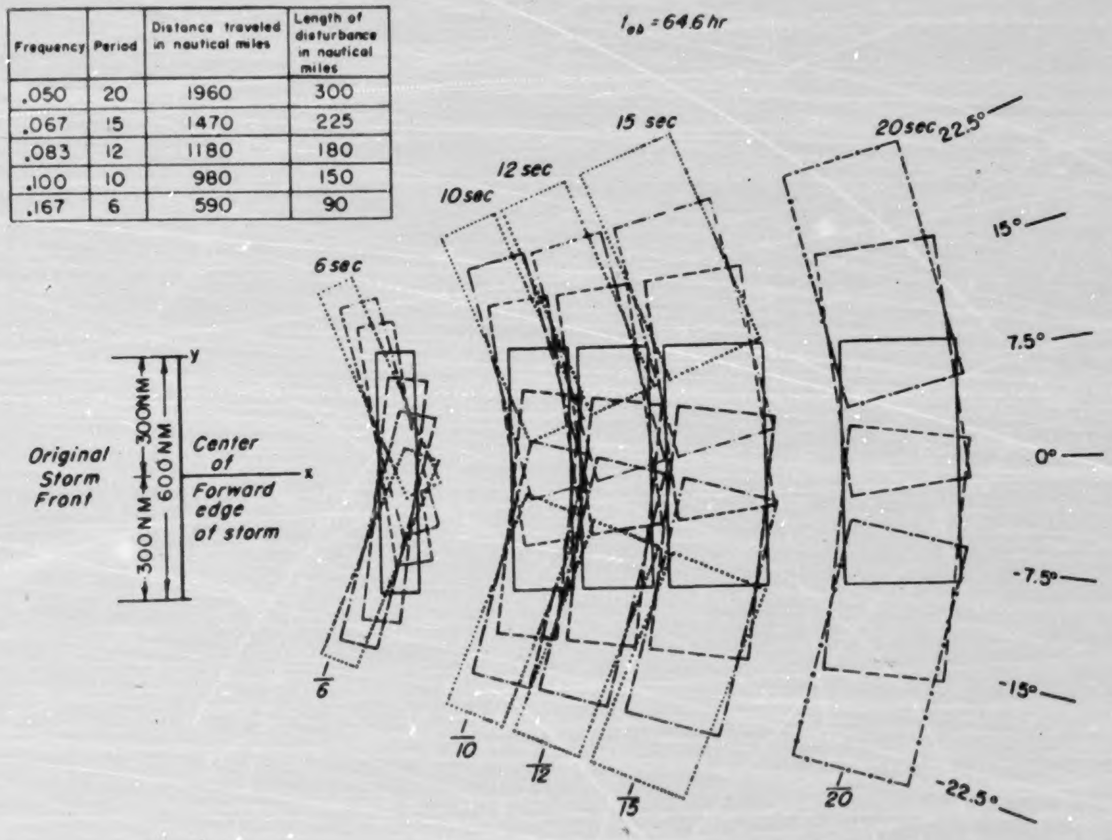

Figure 14-18.—Angular spreading.

continue on, however, the faster runners move ahead and the slower runners begin to fall behind. Thus the field of runners begins to string out along the direction of travel. The wave trains leaving a fetch do the same thing. The stringing out of the various groups of waves is called dispersion.

In a swell forecast problem it is necessary to determine which wave trains have already passed the forecast point and which ones have not yet arrived. After this has been determined, the wave trains which are left are the ones that are at the forecast point at the time of observation.

Angular Spreading

As the wave trains leave the fetch, they may leave at an angle to the main direction of the wind in the fetch. Thus, swell waves may arrive at a forecast point even though it may lie to one side of the main line of direction of the wind. This process of angular spreading is depicted in figure 14-18.

The problem in swell forecasting is to determine how much of the swell will reach the forecast point after the waves have spread out at angles. This is accomplished by measuring the angles from the leeward edge of the fetch to the forecast point. These angles must be measured as accurately as possible and are determined by the following rules. Figure 14-19 illustrates the procedure.

1. Draw the rectangular fetch.

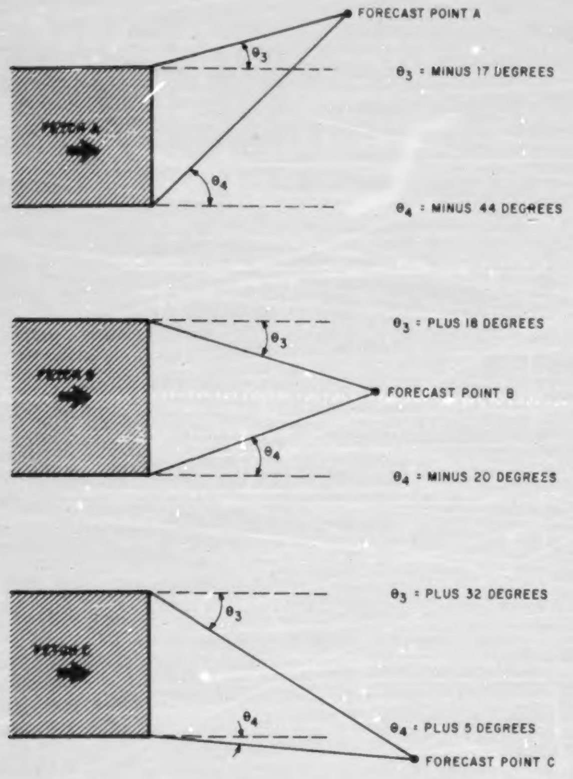

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Figure 14-19.—Measurements of angles for angular spreading.

2. Extend the top and bottom edge of the fetch outward parallel to the main direction of the wind. This is shown as dashed lines in figure 14-19.
3. Draw lines from the top and bottom edges of the fetch to the forecast point.
4. The angles to the forecast point are designated Theta 3 ($\theta 3$) and Theta 4 ($\theta 4$). Theta 3 is measured from the top edge of the fetch and Theta 4 from the bottom edge.
5. Any angle which lies above the dashed line is negative while any angle that lies below the dashed line is positive.

After the angles Theta 3 and 4 have been measured thay are converted to percentages of the swell which will reach the forecast point. This conversion is made by entering sea and swell graph 7, figure 14-20, with the positive or negative angles and reading the corresponding percentages directly. The percentages are then subtracted ignoring the plus or minus to find the angular spreading.

OBJECTIVE METHOD FOR FORECASTING SWELL WAVES

A number of terms used in dealing with forecasting sea waves will be used again in this process; however, a number of new terms will be introduced. Table 14-4 lists most of these terms with their associated symbol and definition.

As with objective forecasting of sea waves a worksheet will be utilized in this procedure. The worksheet is shown in figure 14-21.

The following specific steps for forecasting swell waves coincides with the steps on the worksheet.

STEP 1. Determine the average wind speed over the fetch and enter this value on the worksheet as (U).

STEP 2. Determine how long the wind has maintained the same speed as U by checking back through previous weather charts. Enter the number of hours (t_d) on the worksheet.

STEP 3. Measure the great circle distance from the forecast point to the near edge of the fetch. Enter this distance in miles as the decay distance (D) on the worksheet.

STEP 4. Measure the length of the fetch in nautical miles and enter this value on the worksheet as (F). The measurement must be made parallel to the wind direction over the fetch.

STEP 5. From the sea worksheet determine the upper limit of the period in the fetch (T_u). Record this value on the swell worksheet.

STEP 6. Using the procedure explained previously, measure the angles Theta 3 and 4. Enter sea and swell graph 7 (fig. 14-20), with angles in degrees and read their corresponding percentage values. Record these percentage values on the worksheet.

STEP 7. Subtract the smaller percentage found in Step 6 from the larger. The result is the angular spreading factor (A_s). Record this value on the worksheet.

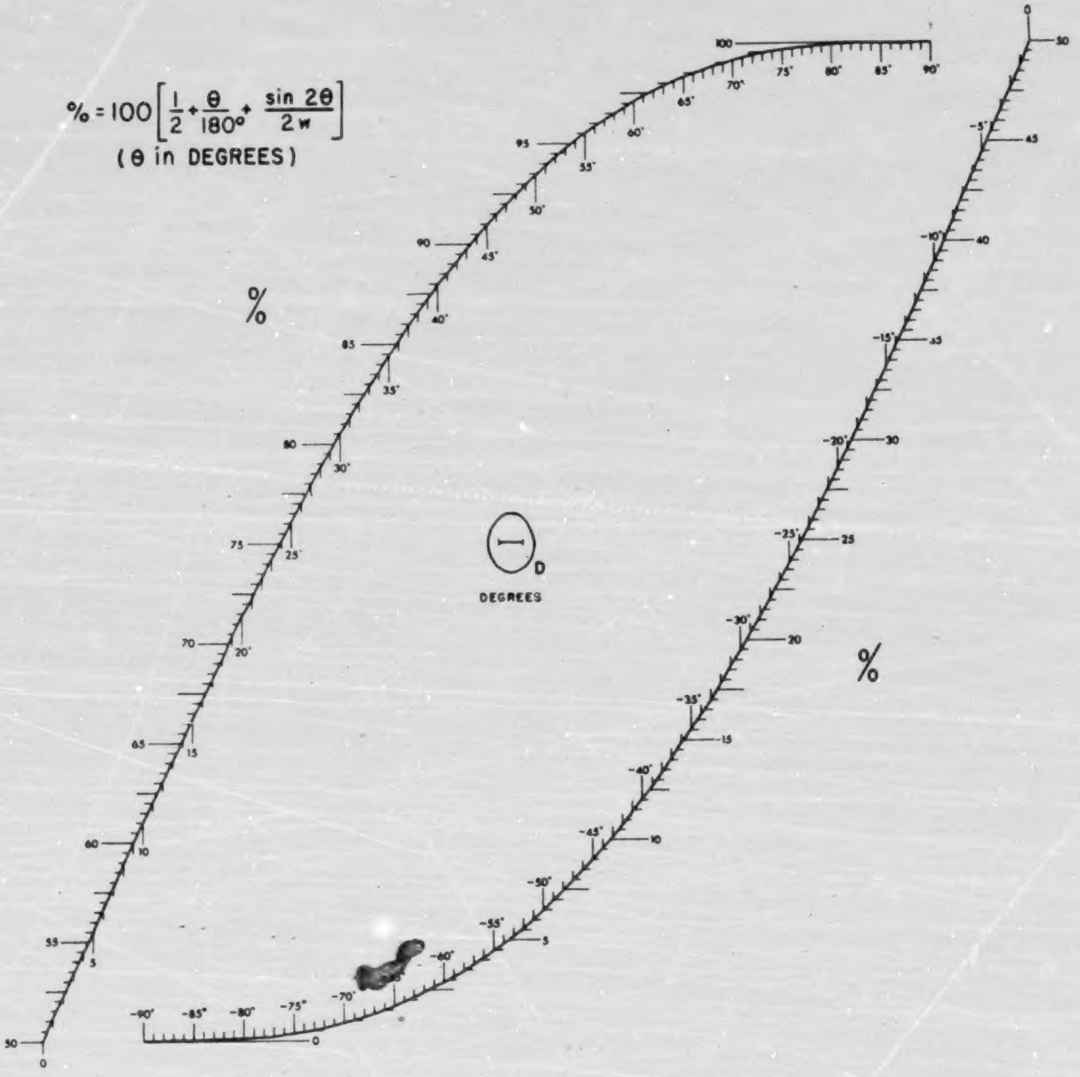

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Figure 14-20.—Sea and swell graph 7.

STEP 8. Turn to sea and swell graph 9, figure 14-22. Enter the left side of the graph with the upper period (T_u) found in Step 5. Enter the bottom of the graph with decay distance (D) found in Step 3. Move to the right from T_u and upward from D and at the intersection read the travel time for the first wave (t_o) in hours on the forecast sheet. This is the time it takes for the first swell wave to arrive at the forecast point.

STEP 9. Add the number of hours for the first wave to arrive to the DTG of the weather chart used to draw the fetch and the result will be the forecast time of arrival of the first wave. Record this value on the worksheet.

STEP 10. Fill in the decay table on the worksheet. The following preliminary discussion is given to provide the forecaster with a better understanding of the use of the decay table.

SWELL WORKSHEET

FETCH #__________ DTG CHART____/________Z DATE__________

* * *

Observed or Forecast Parameters

1. Wind Speed over Fetch: $U =$________kts
2. Wind Duration $t_d =$________hrs
3. Decay Distance $D =$________mi
4. Fetch Length $F =$________mi
5. Upper Limit of Periods in Fetch $T_u =$________sec

* * *

Step No.	Enter Sea & Swell Graph	With	And Read
6.	7	θ_3 and θ_4 measured from map	θ_3 % θ_4 %
7.	None	Subtract the small percentage found in step 6 from the larger	Angular Spreading Factor A_s %
8.	9	T_u from step 5 and D from step 3	Travel time for first wave t_o hrs
9.	None	Add t_o from step 8 to the DTG of the chart for time of arrival of first waves.	ETA____/______Z or ____/______local

10. Fill in Decay Table:

t_{OB}	f_2	T_2	E_2	$t_{OB}-t_d$	f_1	T_1	E_1	ΔE	E_o	H_{avg}	$H_{1/3}$	$H_{1/10}$
t_o												
$t_o + 6$												
$t_o + 12$												
$t_o + 18$												
$t_o + 24$												
$t_o + 30$												
$t_o + 36$												
$t_o + 42$												
$t_o + 48$												

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Figure 14-21.—Swell worksheet.

Table 14-4.—Sea wave terminology.

Name	Symbol and Dimension	Definition
Decay distance	(D) Miles	Distance from point of forecast to the leeward (downwind) edge of the fetch.
Decay index	(H_o/H_F)	Ratio of deep water wave height to fetch wave height.
Deep water wave height	(H_o) Feet	Height of waves after leaving the generating fetch but before reaching shallow water to become surf. (Swell height).
Travel time	(t_o) Hours	Length of time necessary for waves to travel decay distance (D).

Exact instructions for filling in the decay table follow the discussion.

When swell waves reach a forecast point some distance away from the generating area, the waves do not arrive all at once. Instead they gradually build up to a maximum height over a period of hours, then slowly decrease. Because of this, a swell forecast should actually be several forecasts, one for every few hours showing the increasing heights and then the decreasing heights as the swell lowers. The decay table is designed to do this. It is done by forecasting the swell height at the time of arrival of the first wave, then the swell heights for every 6 hours thereafter for the next 48 hours. The time of each forecast is in the t_{ob} column of the decay table. The actual number in the t_{ob} column is the number of hours after the DTG of the map showing the generating area (fetch).

In order to forecast the height of the swell at a given t_{ob} time, it is necessary to know how much of the wave energy has already passed the forecast point and how much has not arrived. That energy which is left is the energy at the forecast point and from this the wave heights can be known. The decay table does this by computing the frequency, period, and energy values just passing the forecast point (f_1, T_1, E_1); computing the frequency, period, and energy values just arriving at the forecast point (f_2, T_2, E_2); and subtracting the arriving energy from the leaving energy. The result is the energy left at the forecast point if there is no angular spreading. To account for angular spreading, the angular spreading factor A_s is multiplied by the remaining energy at the forecast point and from this the wave heights are computed for the given time.

Steps for filling in the decay table are given for each column as follows:

1. t_{ob} Column. Write the value of the travel time of the first wave (t_o) (found in Step 8 on worksheet) as the first number in the t_{ob} column. Each succeeding number down the column should be 6 hours larger than the number before it.
2. f_2 Column. Turn to Sea and Swell Graph 10, figure 14-23. Enter the left side of the graph with the adjacent value of the t_{ob}. Move horizontally to the right and intersect a slanted line of value equal to the decay distance (D), found in Step 3 on the worksheet. From that intersection move downward to the bottom of the graph and read the value of f_2. Repeat this process for each value of t_{ob}.
3. T_2 Column. Turn to sea and swell graph 4, figure 14-14. Enter the left side of the graph with the adjacent value f_2. Move horizontally right to the curved line, then vertically downward to the bottom scale. Read the value of T_2. Repeat this process for each value in the f_2 column.

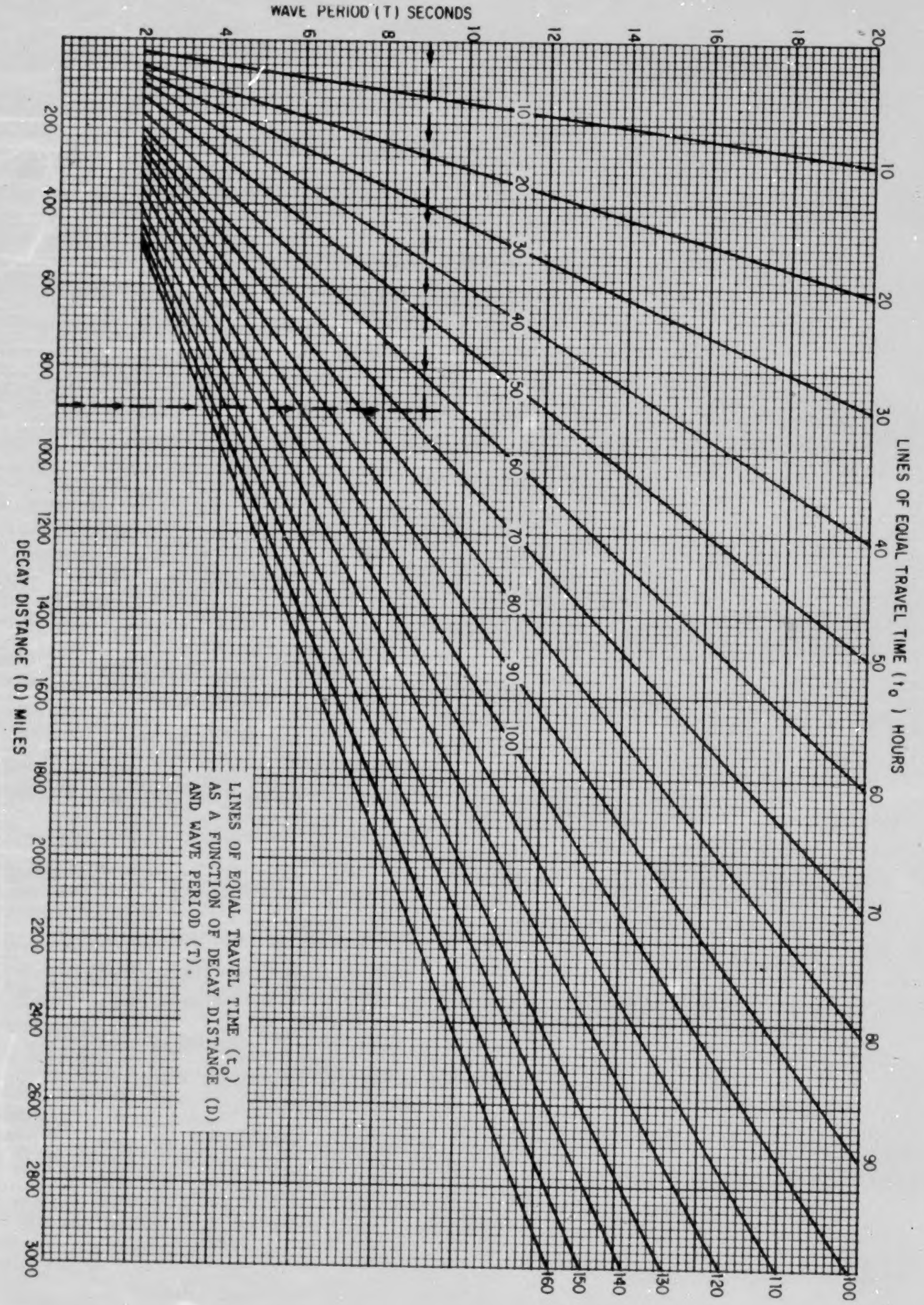

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Figure 14-22.—Sea and swell graph 9.

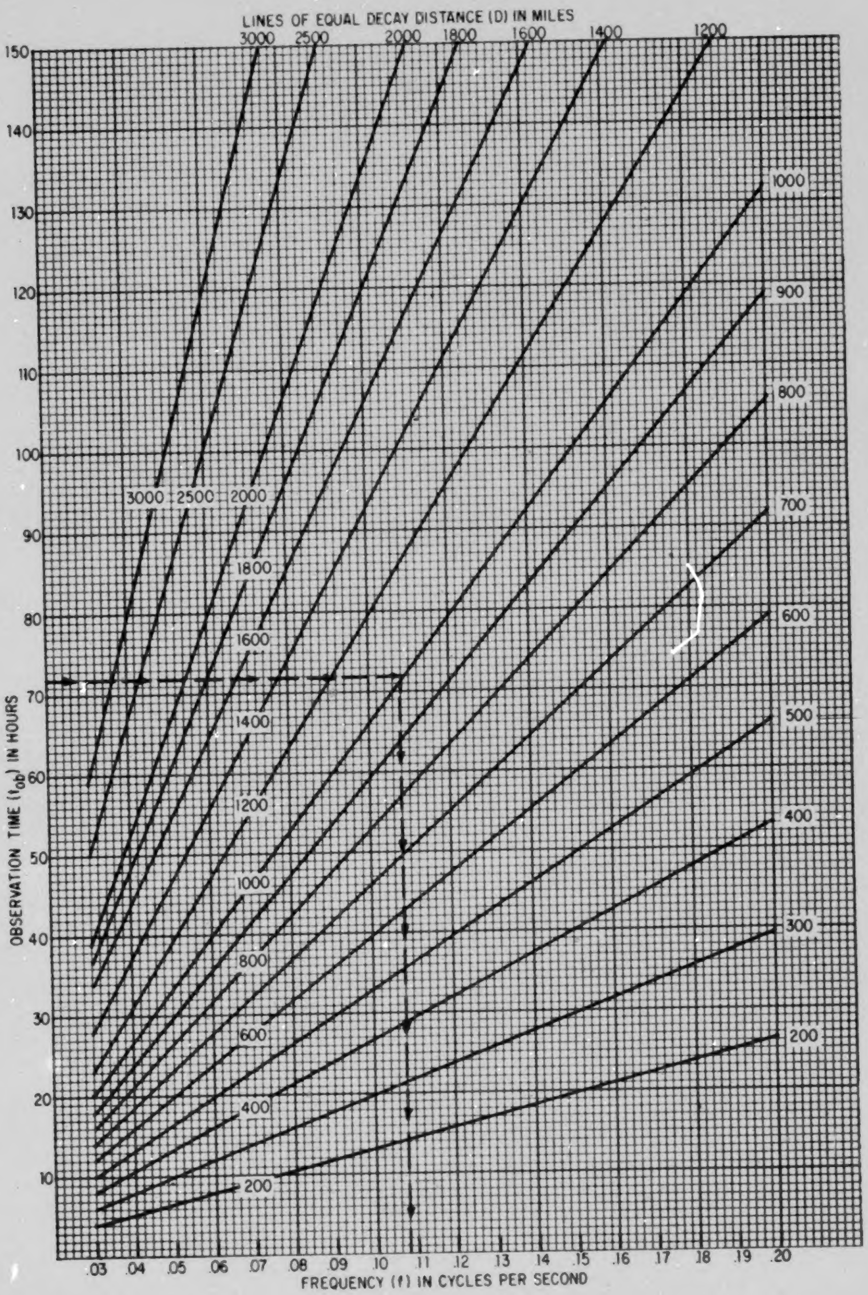

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Figure 14-23.–Sea and swell graph 10.

4. E_2 Column. Turn to any of the C. C. S. graphs (figures 14-7, 14-8, 14-9, or 14-10) having the value of the wind speed (U) from Step 1. Enter the f scale at the bottom of the graph with the adjacent value of f_2 from the f_2 column. Go vertically upward to the wind speed line (U). From this point move horizontally to the left and read E_2 from the E scale. Repeat this process for each value of f_2.

5. t_{ob}-t_d Column. The value of the wind duration was written in Step 2 of the worksheet. Subtract t_d from each adjacent value of t_{ob}. Write the result of each of these subtractions in the t_{ob}-t_d column.

EXCEPTION: If the fetch has been moving to the leeward at nearly the same speed as the wind, and then the wind suddenly stops, then the t_{ob}-t_d column becomes a repeat of the t_{ob} column without subtracting t_d.

6. f_1 Column. Turn again to sea and swell graph 10 (fig. 14-23). Enter the left side of the graph with the number found in the adjacent t_{ob}-t_d column. Move horizontally to the right and intersect the slanted line marked with the decay distance (D) in miles found in Step 3. From that intersection move downward to the bottom of the graph and read the value f. Repeat this process for each value in the t_{ob}-t_d column.

EXCEPTION: If the sea in the fetch is fully developed or if the fetch has been moving nearly as fast as the wind and then suddenly stops then add the fetch length (F) to the decay distance (D) before entering the graph with D.

7. T_1 Column. Turn again to sea and swell graph 4 (fig. 14-14). Enter the left side of the graph with the adjacent value of f_1. Move horizontally right to the curved line, then vertically downward to the bottom scale. Read the value of T_1. Repeat this procedure for each value in the f_1 column.

8. E_1 Column. Turn to any of the C. C. S. graphs used in the computations for the E_2 column, having the correct wind speed. Enter the f scale at the bottom with the adjacent f_1 value. Go vertically upward to the wind speed line (U). From this point move horizontally to the left and read E_1 from the E scale. Repeat this process for each value of f_1.

9. ΔE Column. Subtract the adjacent value of E_2 from E_1. Record this result as ΔE. Repeat this process for each value of E_2 and the adjacent E_1.

10. E_o Column. The angular spreading factor (A_s) was written in step 7 on the worksheet. Multiply this value times the adjacent value of E found in the E column. Record the result as E_o.

11. Characteristic Heights Column. Turn to sea and swell graph 6a or 6b (fig. 14-16 or 14-17). Graph 6a is for E values of 0 to 400 feet squared while 6b is for E values of 0 to 10 feet squared, to allow the smaller values to be read accurately. Select the proper graph for the values of E in the E_o column. Follow the directions at the top of the graph and record the values of the characteristic wave heights.

By completing all the presented steps the forecaster now has all the information necessary to provide an accurate swell forecast.

FORECASTING SURF

Thus far we have discussed the generation of sea waves, their transformation to swell waves, some of the charges that occur as they move, and objective methods of forecasting both waves.

The Navy is also greatly involved in amphibious operations which require the forecasting of another sea surface phenomena, surf. Senior Aerographer's Mates will, upon occasion, provide forecasts for this type of operation. Accurate and timely forecasts can greatly decrease the chances of personnel injuries or equipment damage. It is therefore important that forecasters have a thorough understanding of the characteristics of surf and a knowledge of forecasting techniques.

GENERATION OF SURF

The breaking of waves, in either single or multiple lines along the beach or over some submerged bank or reef is referred to as surf.

The energy which is being expanded in producing this phenomenon is the remainder of that energy that was imparted to the sea surface when the wind developed the sea waves. It, of course, has been depleted as the swell waves transited from the fetch area to the area of occurrence of the surf.

The surf zone is the extent from the water up-rush on the shore to the most seaward breaker. It will be within this area that the forecast will be prepared.

When waves enter an area where the depth of the bottom reaches half their wave length, the waves are said to "feet bottom." This means that the wave is no longer traveling through the water unaltered, but is entering intermediate water where changes in wave length, speed, direction, and energy will occur. There will be no change in period. These changes are known as shoaling and refraction. Shoaling affects the height of the waves, but not direction, while refraction effects both; both effects result from a change in wave speed in shallow water.

Shoaling

The shoaling effect is caused by two factors. The first is a result of the shortening of the wave length; as the wave slows down the crests move closer together. Since the energy between crests remains constant the wave height must increase if this energy is to be carried in a shorter length of water surface. Thus, waves can become higher near shore than they were in deep water. This is particularly true with swell since it has a long wave length in deep water and travels fast. As the swell speed decreases when approaching a shore, the wave length shortens, and a long swell which was barely perceptible in deep water may reach a height of several feet in shallow water. The second factor in shoaling has an opposite effect and is due to the slowing down of the wave velocity until it reaches the group velocity. As the group velocity represents the speed with which the energy of the waves is moving, the height of the individual waves will decrease with its decreasing speed until the wave and group velocities are equal. The second factor predominates when the wave first feels bottom, decreasing the wave height to about 90 percent of its deep water height by the time the depth is one-sixth of the wave length. Beyond that point, the effect of the decreased distances between crests dominates so that the wave height increases to quite large values close to shore.

Refraction

When waves arrive from a direction that is perpendicular to a straight beach, the wave crests will parallel the beach. If the waves are arriving from a direction other than perpendicular or the beach is not straight, the waves will bend, trying to conform to the bottom contours. This bending of the waves is known as refraction and results from the inshore portion of the wave having a slower speed than the portion still in deep water. This refraction will cause a change in both height and direction in shallow water.

Surf Development

When a wave enters water which is shallower than half its wave length, the motion of the water near the bottom is retarded by friction. This causes the bottom of the wave to slow down. As the water becomes more shallow the wave speed decreases, the wave length becomes shorter, and the wave crest increases in height. This continues until the crest of the wave becomes too high and is moving too fast. At this point the crest of the wave becomes unstable and crashes down into the preceding wave trough; when this happens the wave is said to be breaking. The type of breaker, that is, whether spilling, plunging, or surging, is determined by the steepness of the wave in deep water and the slope of the beach.

A spilling breaker breaks gradually over a distance, white water forming at the crest and expanding down the face of the breaker. The wave energy is expanded gradually and breaking is mild.

In a plunging breaker the wave crest advances faster than the base of the wave, causing a wave crest to curl over and break with a crash. The resulting white water appears almost instantly over the complete front face.

Illustrations of both of these type breakers are found in the Rate Training Manual, Aerographer's Mate 3 & 2, NAVTRA 10363-D.

A surging breaker is one that peaks up, but instead of plunging or spilling, surges up on the beach.

When a wave train strikes a beach at an angle refraction occurs, but at the same time another result is the mass transport of water parallel to the beach in the same direction as the movement of the wave train. This mass transport is called the longshore current or Littoral Current. It is of vital importance to amphibious operations.

Definition of Terms

The following are some terms that will be used extensively in surf discussions and that should be understood by the forecaster:

1. Breaker height – the vertical distance in feet between the crest of the breaker and the level of the trough ahead of the breaker.
2. Breaker wave length – the horizontal distance in feet between successive breakers.
3. Breaker period – the time in seconds between successive breakers. This is always the same as the deepwater wave period.
4. Depth of breaking – the depth of the water in feet at the point of breaking.
5. Surf zone – the horizontal distance in yards between the outermost breakers and the limit of wave uprush on the beach.
6. Number of lines of surf – the number of lines of breakers in the surf zone.
7. Deep water wave angle – the angle between the bottom contours and the deep water swell wave crests.
8. Breaker angle – the angle between the beach and the lines of breakers. It is always less than the deep water wave angle.

OBJECTIVE TECHNIQUE FOR FORECASTING SURF

The objective technique presented in this manual is the same method contained in NWRF 36-1264-099, Surf Forecasting. Some minor innovations in the graphs used were made by personnel of the Oceanographic Department of Naval Weather Service Facility, San Diego.

Through the use of the procedure as presented in this chapter the forecaster can provide an accurate forecast of surf conditions when required.

Figure 14-24 provides an example of the surf worksheet that may be utilized in this surf forecasting procedure. The steps in the method conform to steps on the worksheet.

STEP 1. Determine the deep water wave height approaching the beach from either observations or a swell forecast. Enter this value as H_o on the worksheet.

STEP 2. Determine the deep water wave period approaching the beach from either observations or a swell forecast. Enter this value as T_o on the worksheet.

STEP 3. From the weather chart determine the direction from which the swell waves will approach the beach. Assuming that the line along the wave crests will be perpendicular to be direction of wave travel, the angle between the beach contours and the deep water wave crests can be measured. Enter this value as a_o on the worksheet.

STEP 4. Turn to surf graph 1, figure 14-25. Enter the left side with H_o found in step 1, and the bottom with T_o from step 2. At the intersection of these values read $H_o/T_o{}^2$ from the curved solid lines of the graph. Enter this value on the worksheet.

STEP 5. Turn to surf graph 2, figure 14-26. Enter the bottom with $H_o/T_o{}^2$ found in step 4 and go up vertically to the curved line marked with the beach slope of the beach for which the forecast is being made. From that point go to the left horizontally and read the value of H_b/H_o at the left side of the graph. Record this value on the worksheet.

STEP 6. Turn to surf graph 3, figure 14-27. Enter the left side with H_o found in step 1 and the bottom with H_b/H_o from step 5. At the point of intersection read the value of H_b from the curved lines on the graph. Enter this value on the worksheet.

STEP 7. Turn to surf graph 4, figure 14-28. Enter the bottom with H_o/T_o found in step 4 and go up vertically to the line marked with the slope of the beach for which the forecast is being made. Read the breaker type written at that point and enter the type on the worksheet.

STEP 8. Turn to surf graph 5, figure 14-29. Enter the bottom with $H_o/T_o{}^2$ found in step 4. Go up vertically to the curved line on the graph and from that point of intersection go across horizontally to the left side. At the left side of the graph read the value of d_b/H_o and record it on the worksheet. If $H_o/T_o{}^2$ is less than .01 go directly to step 9.

SURF FORECAST WORKSHEET

BEACH NAME____________________BEACH SLOPE________________FTA SURF__________

****

FROM OBSERVED OR FORECAST SWELL

1. Deep water wave height: H_o =_________ft.
2. Deep water wave period: T_o =_________sec.
3. Angle between deep water waves and depth contours: a_o =_________deg.

****

SURF CALCULATIONS

Step	Enter SURF GRAPH	With	And Read
4	1	H_o from step 1 and T_o from step 2	H_o/T_o^2
5	2	H_o/T_o^2 from step 4	H_b/H_o
6	3	H_o from step 1 and H_b/H_o from step 5	H_b ft.
7	4	H_o/T_o^2 from step 4 and Beach Slope from heading	Breaker Type
8	5	H_o/T_o^2 from step 4. If $H_o/T_o^2 < .01$ go to next step	d_b/H_o
9	6	H_o from step 1 and d_b/H_o from step 8 or use $d_b = 1.3\ H_o$ if $H_o/T_o^2 < .01$	d_b ft.
10	7	d_b from step 9 and Beach Slope from heading	Width of Surf Zone yds.
11	8	d_b from step 9 and T_o from step 2	L_b ft.
12	9	L_b from step 11 and Width of Surf Zone from step 10.	No. Lines of Surf
13	10	d_b from step 9 and T_o from step 2.	d_b/L_o
14	11	a_o from step 3 and d_b/L_o from step 13	a_b deg. K_d
15	12	H_b from step 6 and K_d from step 14	H_b corrected for refraction cor H_b ft.
16	13	a_b from step 14 and Beach Slope from heading. H_b from step 15 and T_o from step 2	Longshore Current kts.

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Figure 14-24.—Sample surf worksheet.

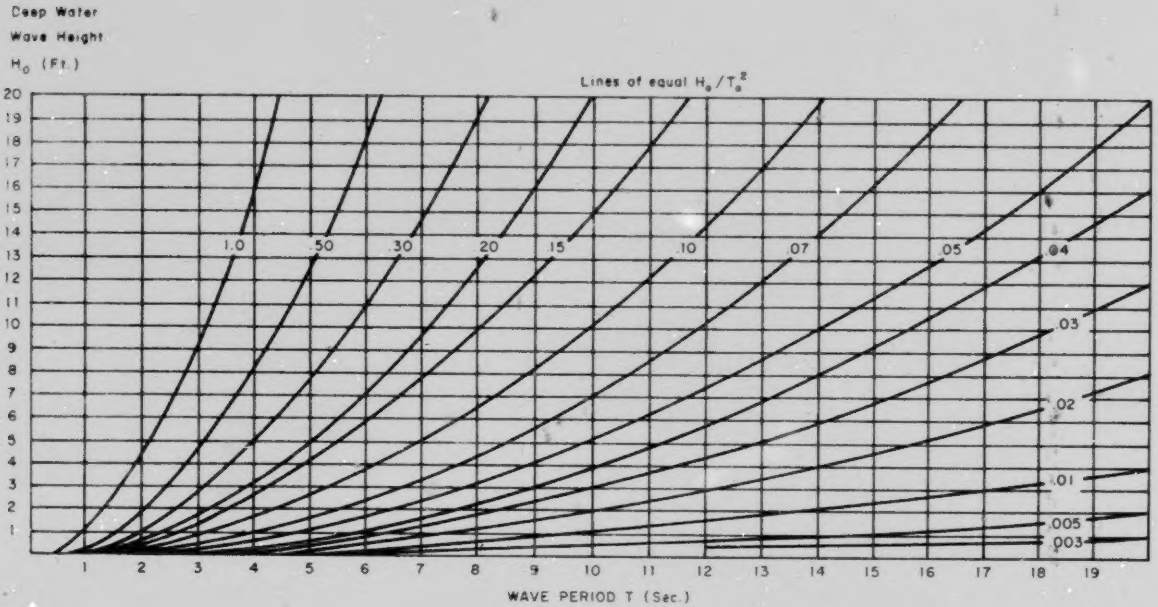

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Figure 14-25.—Surf graph 1.

STEP 9. Turn to surf graph 6, figure 14-30. Enter the side of the graph with H_o found in step 1 and the bottom with d_b/H_o found in step 8. At the intersection read d_b from the curved lines on the graph and record this value on the worksheet. If $H_o/T_o{}^2$ is less than .01 in step 8, do not use surf graph 6. Instead use the formula

$$d_b = 1.3H_b$$

STEP 10. Turn to surf graph 7, figure 14-31. Enter the left side with d_b found in step 9 and go across horizontally to the line marked with the beach slope. From this point go down to the bottom of the graph and read the width of the surf zone. Record this value on the worksheet.

STEP 11. Turn to surf graph 8, figure 14-32. Enter the left side with d_b found in step 9 and go across horizontally to the line marked with the value of T_o found in step 2. From this point go down to the bottom of the graph and read the value of L_b. Record this value on the worksheet.

STEP 12. Turn to surf graph 9, figure 14-33. Enter the left side with the width of the surf zone found in step 10 and enter the bottom with L_b found in step 11. At the intersection read the number of lines of surf. If a whole number of lines is not found, record the value in terms of "1 or 2 lines," etc.

STEP 13. Turn to surf graph 10, figure 14-34. Enter the left side with d_b found in step 9 and the bottom with T_o found in step 2. At the intersection read the value of d_b/L_o from the curved lines. Record this value on the worksheet.

STEP 14. Turn to surf graph 11, figure 14-35. Enter the left side with a_o found in step 3 and the bottom with d_b/L_o found in step 13. At the intersection read a_b from the curved dashed lines of the graph and read K_d from the curved solid lines. Record the values of a_b and K_d on the worksheet.

STEP 15. Turn to surf graph 12, figure 14-36. Enter the left side with H_b found in step 6 and the bottom with K_d found in step 14. At the

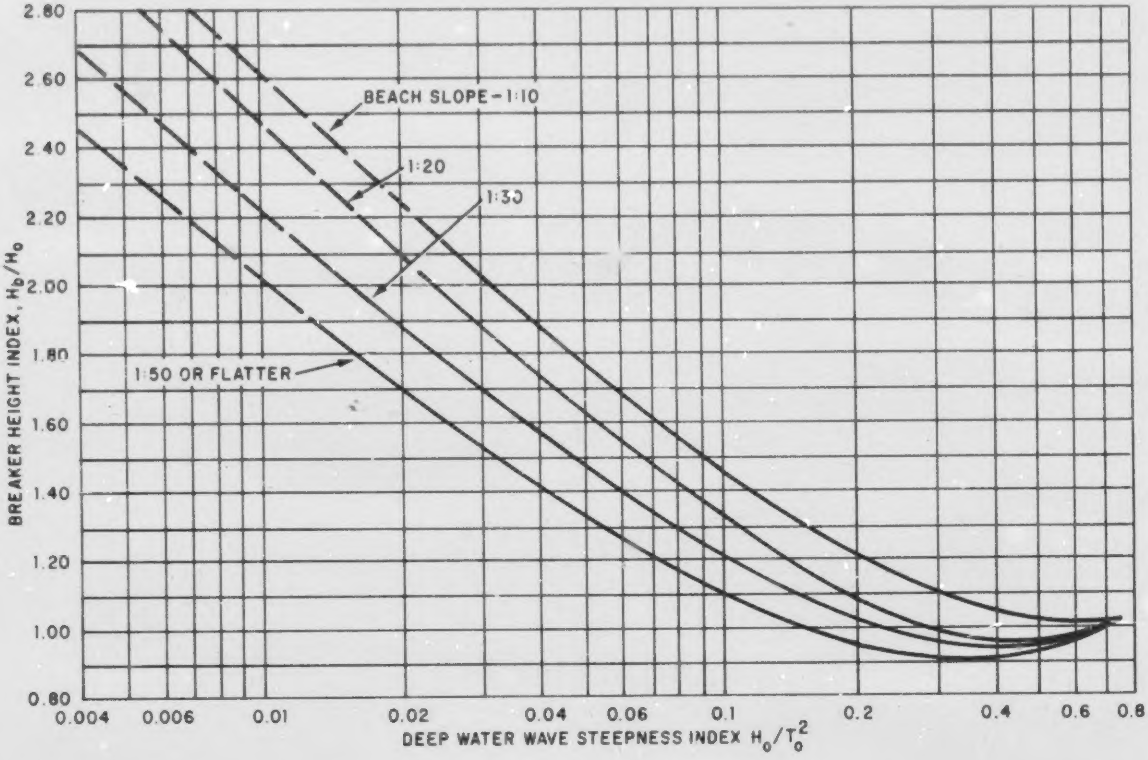

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Figure 14-26.—Surf graph 2.

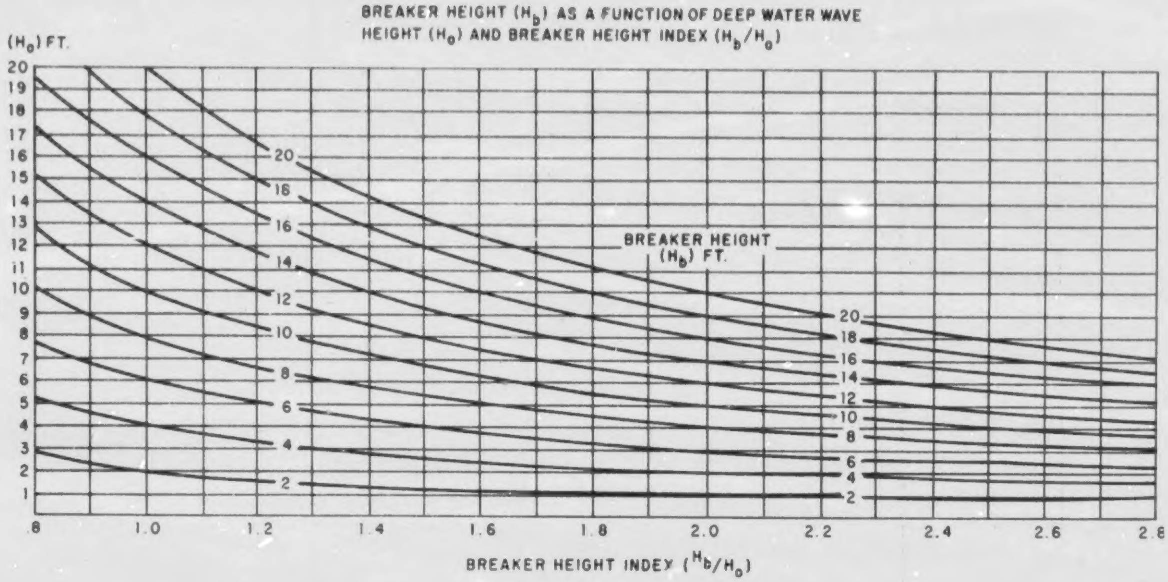

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Figure 14-27.—Surf graph 3.

intersection read a new value for H_b corrected for refraction. Record this new value on the worksheet.

STEP 16. Turn to surf graph 13, figure 14-37. This is a nomogram to determine the longshore current. The instructions for its use are written on it. Record this value on the worksheet.

All of the necessary information to provide a complete surf forecast is now available. The presentation to the user can be made in any manner that is agreed upon; however, figure 14-38 illustrates one of the most commonly used methods that is employed.

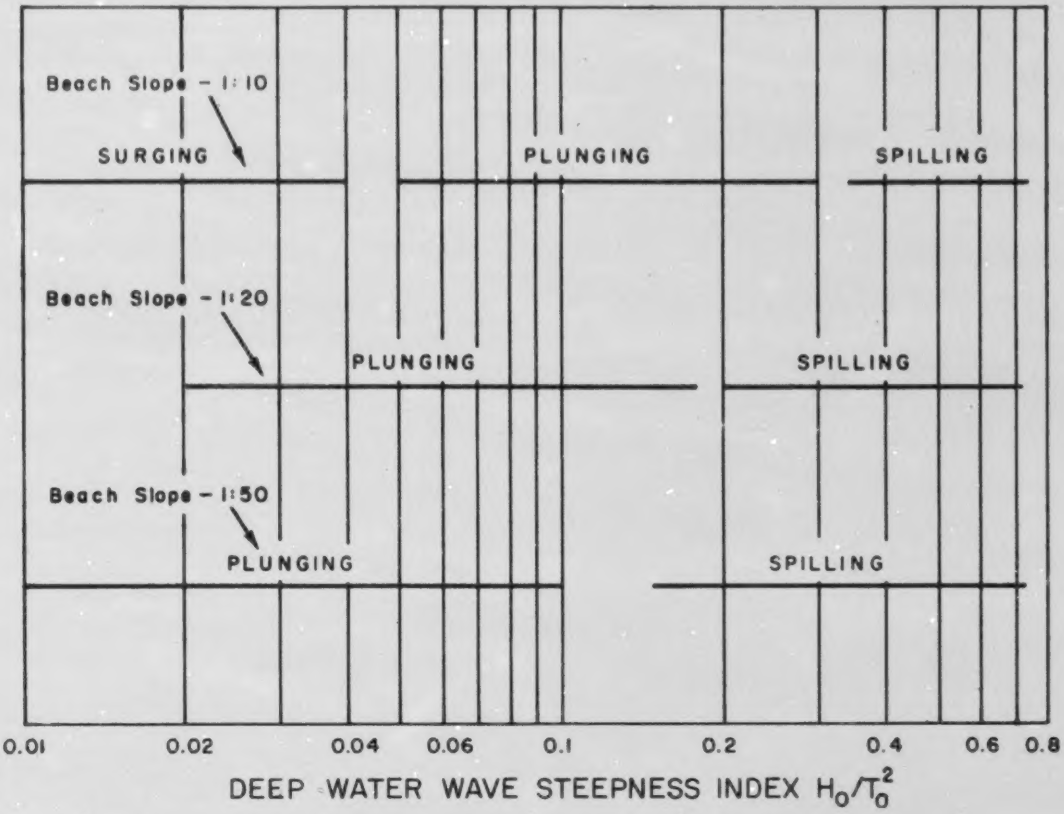

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Figure 14-28.—Surf graph 4.

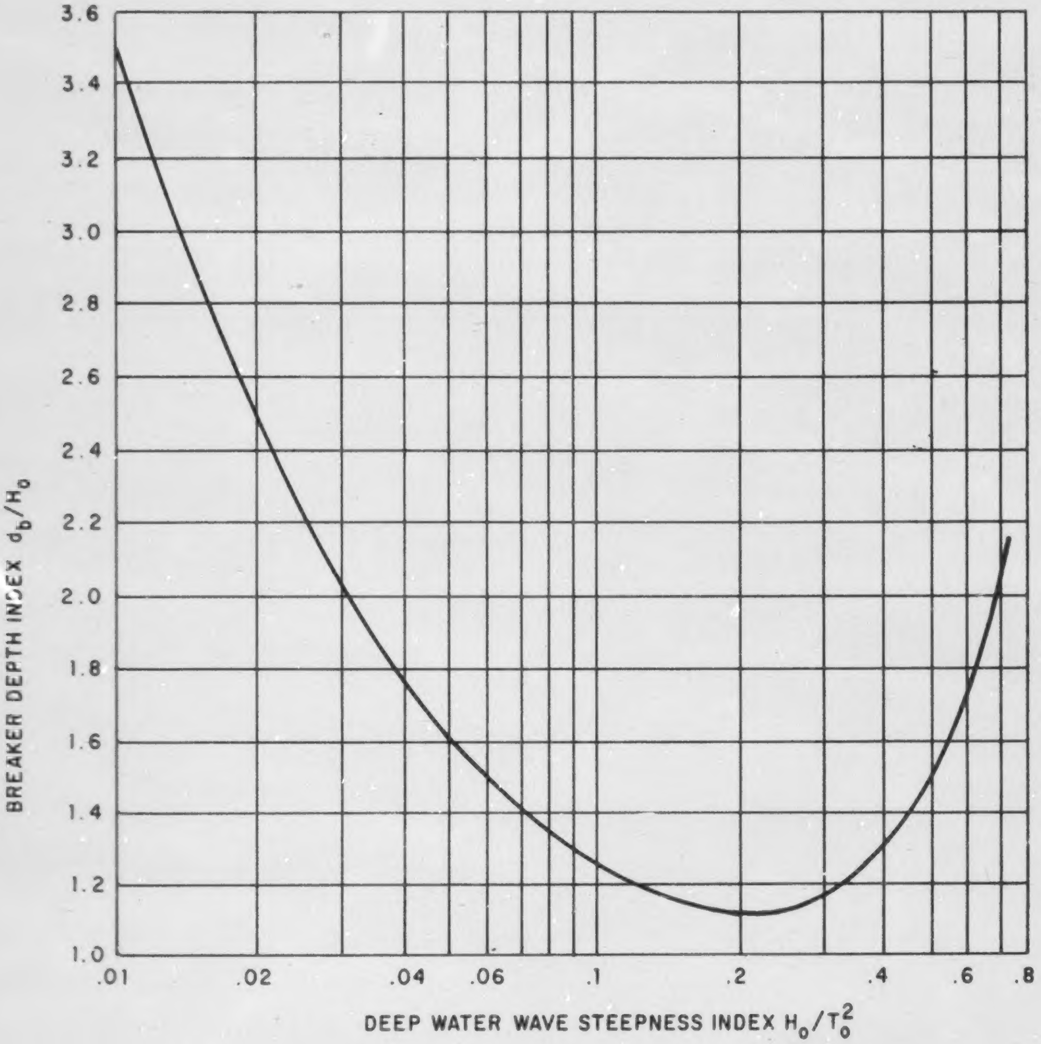

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Figure 14-29.—Surf graph 5.

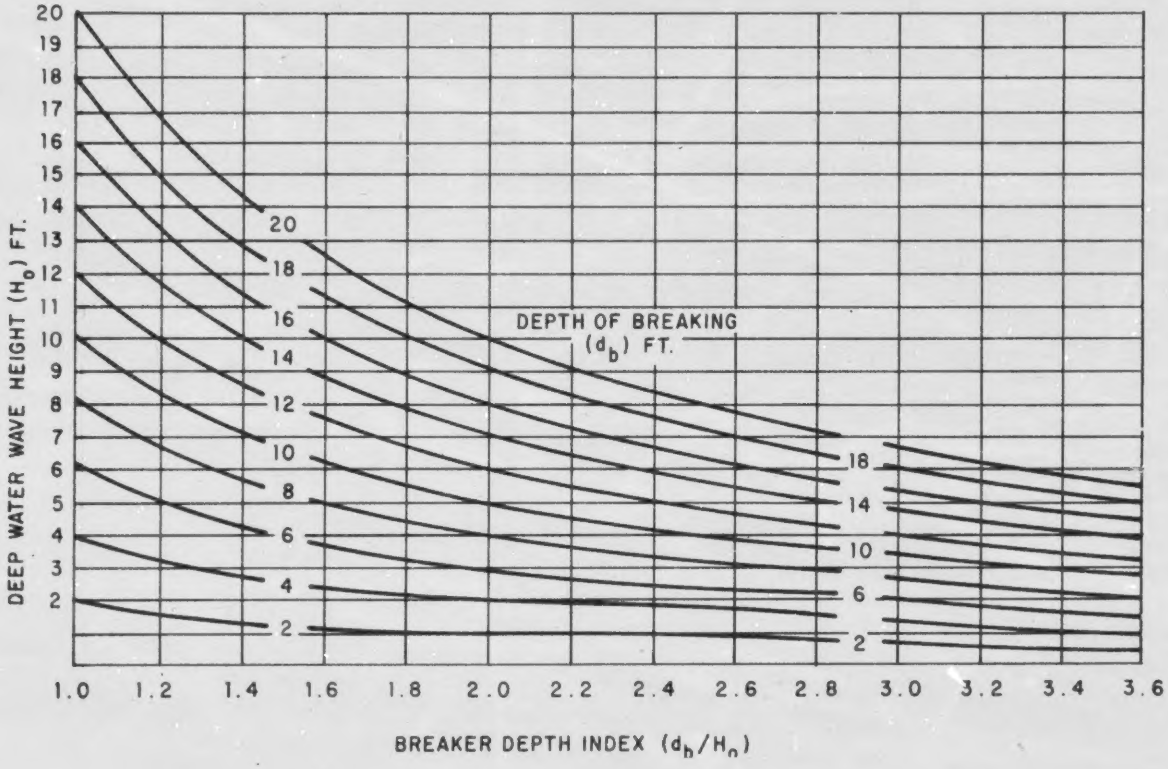

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Figure 14-30.–Surf graph 6.

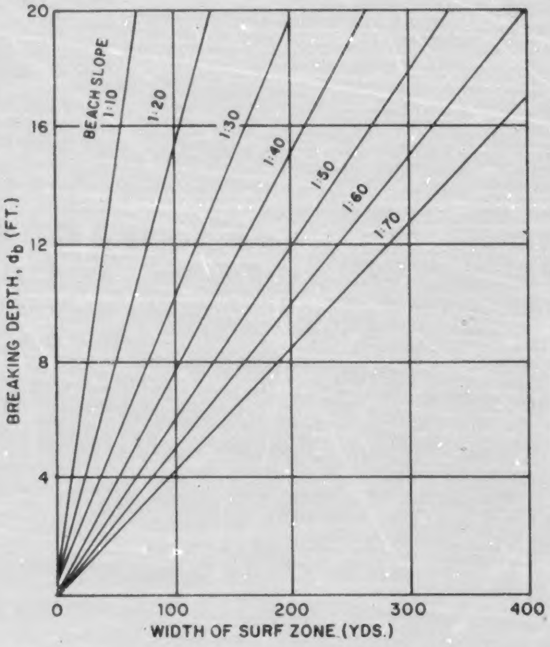

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Figure 14-31.—Surf graph 7.

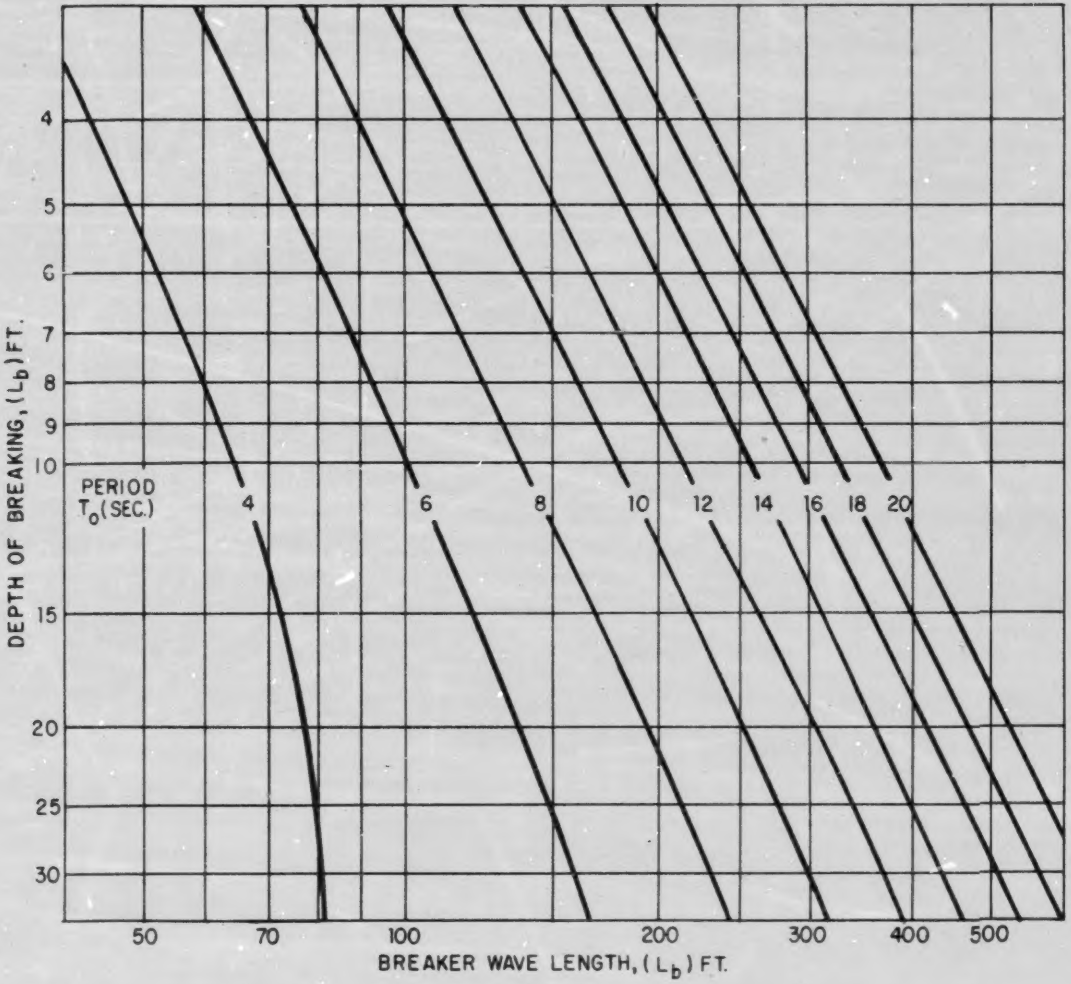

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Figure 14-32.—Surf graph 8.

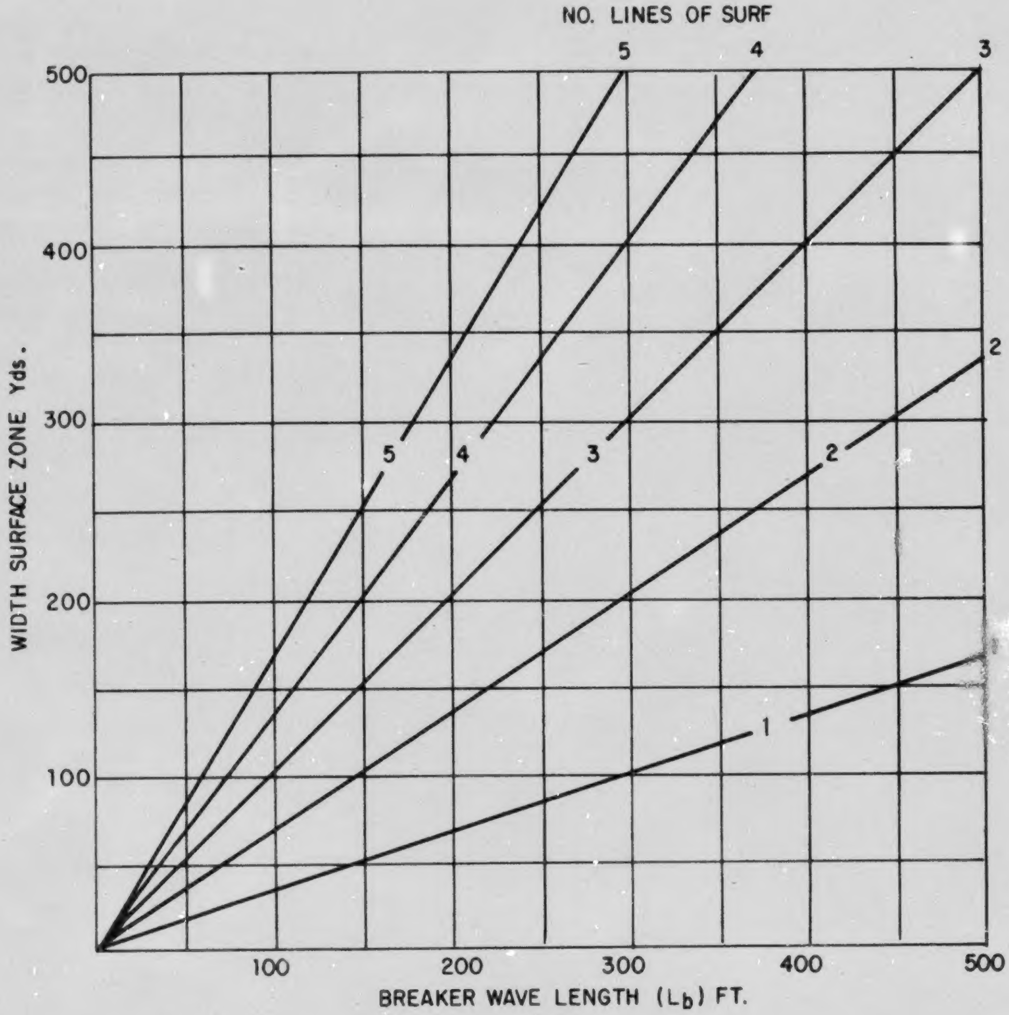

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Figure 14-33.—Surf graph 9.

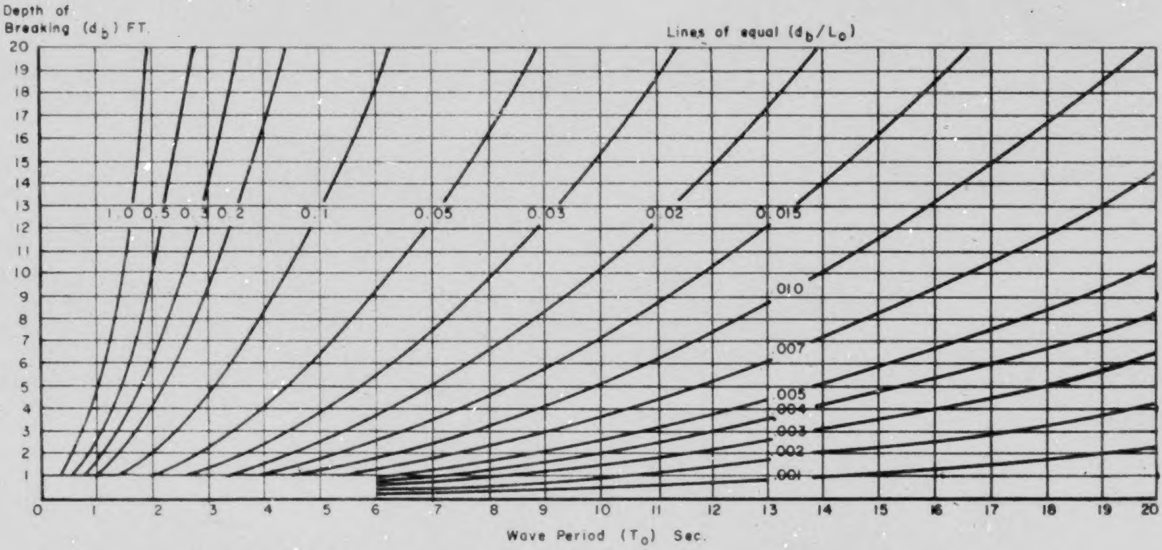

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Figure 14-34.—Surf graph 10.

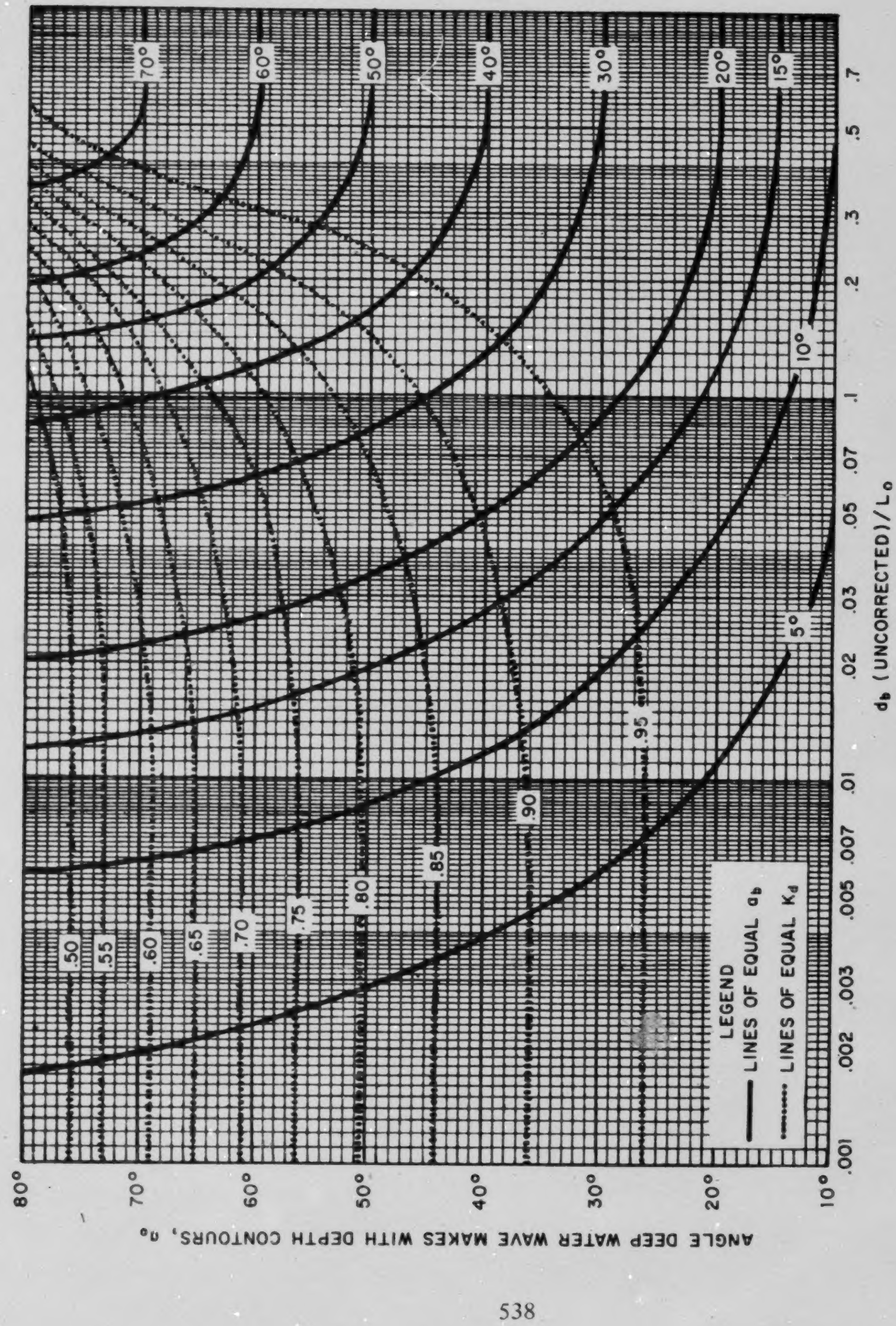

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Figure 14-35.—Surf graph 11.

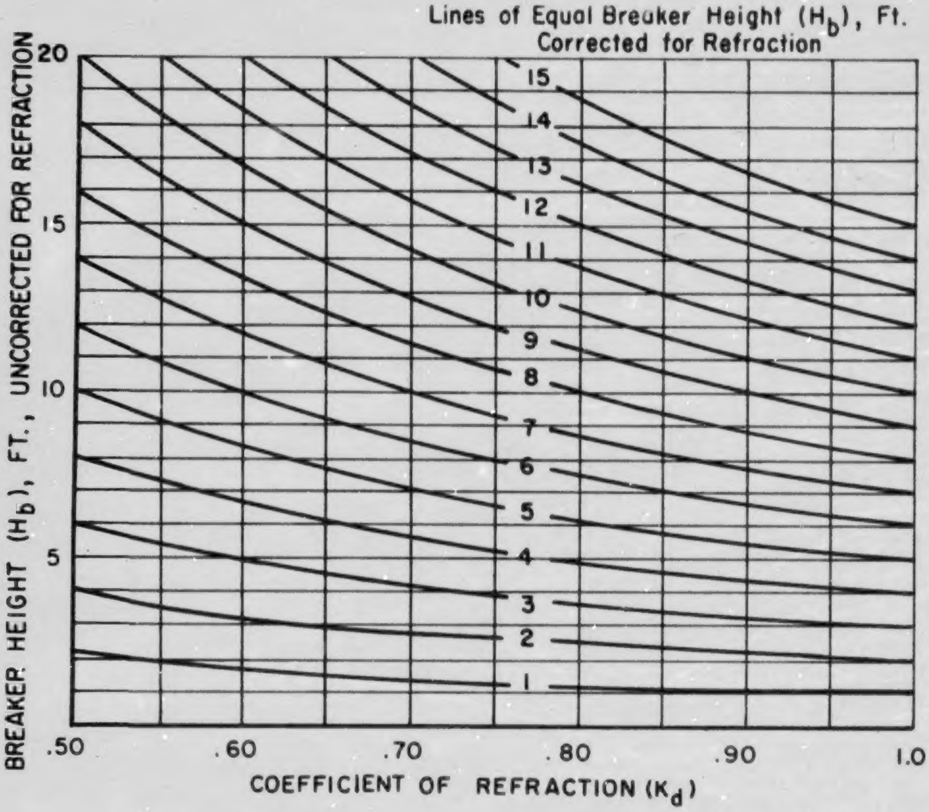

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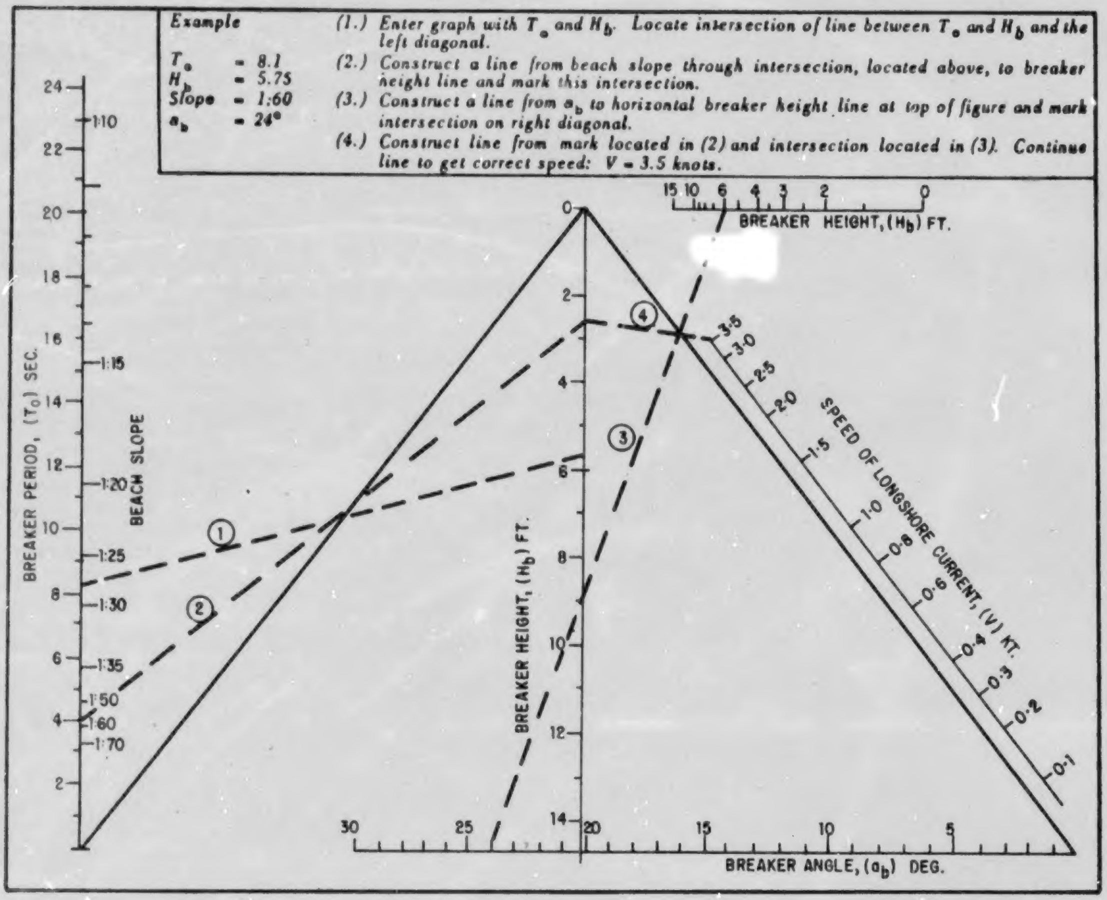
Figure 14-36.—Surf graph 12.



Example

T_o = 8.1
H_b = 5.75
Slope = 1:60
α_b = 24°

(1.) *Enter graph with T_o and H_b. Locate intersection of line between T_o and H_b and the left diagonal.*

(2.) *Construct a line from beach slope through intersection, located above, to breaker height line and mark this intersection.*

(3.) *Construct a line from α_b to horizontal breaker height line at top of figure and mark intersection on right diagonal.*

(4.) *Construct line from mark located in (2) and intersection located in (3). Continue line to get correct speed: V = 3.5 knots.*

BREAKER PERIOD, (T_0) SEC.
24
22
20
18
16
14
12
10
8
6
4
2
BEACH SLOPE
1:10
1:15
1:20
1:25
1:30
1:35
1:50
1:60
1:70
①
②
③
④
BREAKER HEIGHT, (H_b) FT.
0
2
4
6
8
10
12
14
15 10 6 4 3 2 0
BREAKER HEIGHT, (H_b) FT.
SPEED OF LONGSHORE CURRENT, (V) KT.
3.5
3.0
2.5
2.0
1.5
1.0
0.8
0.6
0.4
0.3
0.2
0.1
30
25
20
15
10
5
BREAKER ANGLE, (α_b) DEG.

AG.710

Figure 14-37.—Surf graph 13.

SURFCST ____________ (Beach) ____________ (Time) ____________

ALPHA ____________ BRAVO ____________ CHARLIE ____________

DELTA ____________ ECHO ____________ FOXTROT ____________

HOTEL ____________

ALPHA = Significant Breaker Height (ft.)

BRAVO = Maximum Breaker Height (ft.)

CHARLIE = Period of Breakers (sec.)

DELTA = Type of Breakers

ECHO = Angle Breakers Make with Beach (deg.)

FOXTROT = Longshore Current (kts.)

GOLF = Number of Lines of Surf, Width of Surf Zone (yd.)

HOTEL = Remarks

AG.711

Figure 14-38.—Example of final forecast form.

FORECASTING SURFACE CURRENTS

Although the forecasting of surface currents has been performed by weather service personnel for a number of years, the prominence of such forecasting became more evident when a number of incidents involving large sea-going oil tankers occurred. Collisions and groundings involving tankers caused great amounts of pollutants, oil mainly, to be spilled on the water surface. The movement, both direction and speed, of such contaminants is directly controlled by the surface currents in the affected area. More concerned emphasis has now been placed on the ability of forecasters to predict the movement of such contaminated areas.

In the past, weather service units have provided forecasts to assist in the location of personnel or boats adrift in the open sea as well as forecasts utilized in estimating ice flow.

With the continued growing concern about pollution and contamination of ocean waters, it is anticipated that more requests for current and drift forecasts will be directed to weather service units.

In this section we will discuss the general characteristics of currents, how they form, and different types of currents. There are presently no hard and fast rules or techniques that are universally followed. Most weather units involved in providing such forecasts have their own innovations and methods.

CURRENTS

Aerographer's Mates have a knowledge of the major ocean currents and the meteorological results of the interaction of sea and air. Oceanic circulation (currents) plays a major role in the production of and distribution of weather phenomena. Principle surface current information such as direction, speed, and temperature distribution is relatively well known.

Currents in the sea are generally produced by wind, tide, differences in density between water masses, sea level differences, or runoff from the land. They may be roughly classed as tidal or nontidal currents. Nontidal currents include the permanent currents in the general circulatory systems of the oceans; geopotential currents, which are those associated with density differences in water masses; and temporary currents, such as wind-driven currents which are developed from meteorological conditions. Tidal currents are usually significant in shallow water only, where they often become the strong or dominant flow.

The system of currents in the oceans of the world keeps the water continually circulating. The positions shift only slightly with the seasons except in the Southeast Asia area where monsoonal effects actually reverse the direction of flow from summer to winter. Currents appear on most charts as well behaved continuous streams defined by clear boundaries and with gradually changing directions. These presentations usually are smoothed patterns which were derived from averages of many observations.

The speed of a current is known as its drift. Drift is normally measured in knots. The term velocity is often interchanged with the term speed in dealing with currents although there is a difference in actual meaning. Set, the direction in which the current acts or proceeds, is measured according to compass points or degrees. Observations of currents are made directly by mechanical devices that record speed and direction, or indirectly by water density computations, drift bottles, or visually using slicks and water color differences.

Ocean currents are usually strongest near the surface and sometimes attain considerable speed, such as 5 knots or more reached by the Florida Current. In the middle latitudes, however, the strongest surface currents rarely reach speeds above 2 knots.

Eddies, which vary in size from a few miles or more in diameter to 75 miles or more in diameter, branch from the major currents. Large eddies are common on both sides of the Gulf Stream from Cape Hatteras to the Grand Banks. How long such eddies persist and retain their characteristics near the surface is not well known but large eddies near the Gulf Stream are known to persist longer than a month. The surface speeds of currents within these eddies, when first formed, may reach 2 knots. Smaller eddies have much less momentum and soon die down or lose their surface characteristics through wind stirring.

Wind Driven Currents

Wind driven currents are, as the name implies, currents that are created by the force of the wind exerting stress on the sea surface. This stress causes the surface water to move and this movement is transmitted to the underlying water to a depth which is dependent mainly on the strength and persistance of the wind. Most ocean currents are the result of winds that tend to blow in a given direction over considerable amounts of time. Likewise, local currents, those peculiar to an area in which they are found, will arise when the wind blows in one direction for some time. In many cases the strength of the wind may be used as a rule of thumb for determining the speed of the local current; the speed is figured as 2 percent of the wind's force. Therefore, if a wind blows 3 or 4 days in a given direction at about 20 knots, it may be expected that a local current of nearly 0.4 knot is being experienced.

A wind-driven current does not flow in exactly the same direction as the wind, but is deflected by the earth's rotation. The deflecting force (Coriolis force) is greater at high latitudes and more effective in deep water. It is to the right of the wind direction in the Northern Hemisphere and to the left in the Southern Hemisphere. At latitudes between 10N and 10S the current usually sets downwind. In general the angular difference in direction between the wind and the surface current varies from about 10 degrees in shallow coastal areas to as much as 45 degrees in some open ocean areas. The angle increases with the depth of the current and at certain depths the current may flow in the opposite direction to that of the surface.

Some major wind-driven currents are the West Wind Drift in the Antarctic, the North and South Equatorial Currents that lie in the trade wind belts of the ocean, and the seasonal monsoon currents of the Western Pacific.

Coastal and Tidal Currents

Coastal currents are caused mainly by river discharge, tide, and wind. However, they may in part be produced by the circulation in the open ocean areas. Because of tides or local topography, coastal currents are generally irregular.

Tidal currents, a factor of little importance in general deepwater circulation, are of great influence in coastal waters. The tides furnish energy through tidal currents, which keep coastal waters relatively well stirred. Tidal currents are most pronounced in the entrances to large tidal basins that have restricted openings to the sea. This fact often accounts for steerage problems experienced by vessels.

WIND DRIVEN CURRENT PREDICTION

Attempts at current prediction in the past have only been moderately successful. There has been a tendency to consider ocean currents in much the same manner as wind currents in the atmosphere, when in actuality it appears that ocean currents are affected by an even greater number of factors. It therefore requires different techniques to be utilized.

In order to predict current information it must be understood that currents are typically unsteady in direction and speed. This has been well documented by a number of studies that have been conducted. The reasons for this variability has been attributed to the other forces besides wind and tides which affect the currents.

Climatological surface charts have been constructed for nearly all the oceans of the world using data from ship's drifts. However, this data has been shown to have limitations and should be used as a rough estimate only.

Synoptic Analysis and Forecasting of Surface Currents, NWRF 36-0667-127, provides a composite method of arriving at current forecasts. This method utilizes portions of other methods that have been used. Forecasters should make themselves aware of the information contained in this publication.

COASTAL AND TIDAL CURRENT PREDICTION

Prediction of tidal currents must be based on specific information for the locality in question. Such information is contained in various forms in many navigational publications.

Tidal Current Tables, issued annually, list daily predictions of the times and strengths of flood and ebb currents and the time of intervening slacks. Due to lack of observational data, coverage is considerably more limited than for tides. The Tidal Current Tables do include supplemental data by which tidal currents can be determined for many places in addition to those for which daily predictions are given.

NUMERICAL SEA SURFACE ANALYSES AND FORECASTS

Analyses of a number of sea surface elements is carried out on a routine basis. Although these charts are not readily available as analyses they are utilized in the preparation of prognostic charts which are disseminated.

Sea surface temperature charts, wave height, and wind drift currents are examples of the type of charts available on the fleet facsimile broadcast. Up to date schedules should reflect the charts currently available as this is subject to change. Tailored forecasts for sea conditions are available on a request basis from weather facilities and centrals. Oceanographic departments within these organizations are prepared to provide this service.

Ratt-graphic types of messages are also utilized to disseminate sea condition information over the fleet teletype broadcast.

CHAPTER 15

OCEAN THERMAL STRUCTURE FORECASTING AND ASWEPS

It is generally agreed that presently the submarine poses the greatest military threat to the security of the United States. They may be employed to fire missiles at inland targets as well as disrupt the merchant shipping which is so critical to our nation's survival. Since World War II the Navy has directed a concerted effort to the improvement of methods and equipment to detect these undersea weapons.

The most effectual means of reducing their effectiveness is by detecting them within their own environment, under the ocean surface. Sound is employed in different manners to accomplish this; the active and passive detection systems being examples. Active systems put sound into the water with the echo returning from the submarine, being sensed by the detector. The passive system utilizes sound sensing equipment to detect the noise emitted by the submarine. Surface ships and aircraft, including helicopters, employ these systems or variations of them.

Most people are aware that sound is affected to a great extent by the changes in air temperature. The temperature of the water also plays games with the sound waves as they pass through it.

Aerographer's Mates work in close association with units involved in ASW operations. It is important that they have a thorough understanding of the thermal structure of the ocean, as well as what numerical products are available, their application, and what type of services and assistance are provided them by support activities. These areas will be discussed in this chapter.

In chapter 22 of Aerographer's Mate 3 & 2, NavTra 10363-D, illustrations and definitions of the basic terminology is presented. It is recommended that personnel review this chapter for application to the data as presented in this manual.

THERMAL STRUCTURE FORECASTING

Bathythermograms show that the ocean is more or less stratified. Two points separated by several hundred yards but at the same depth will have practically the same temperature. If the ocean were in equilibrium, this stratification would be complete; the warm lighter water being at the surface, the lower strata consisting of cooler, heavier water, and the boundaries between strata being horizontal surfaces. This equilibrium is disturbed by three processes: advection, the heat budget, and mixing.

OCEAN THERMAL EFFECTS ON ECHO RANGING

The direction that a sound wave will travel in the ocean is largely dependent upon the speed of the individual sound wave or ray within the beam. The speed of sound in sea water depends on the properties of the water. In general these properties vary both horizontally and vertically, but only temperature is of any great significance. Changes in velocity with depth, even slight ones due to warming of the surface water on a bright calm day, deflect the sound beam from the desired straight path and may cause it to overshoot or undershoot an object.

The velocity of sound is directly proportional to the temperature of the medium. The speed of sound in air is 331.5 meters per second at 0 degrees Celsius and increases 0.6 meters per second for each degree increase in temperature. In sea water with salinity of 35 parts per thousand at 0 degrees Celsius the speed of sound is 1449.1 meters per second or 4.4 times the speed of sound in air. The rate at which speed increases with an increase in temperature is not uniform; it is greater at lower temperatures. Figure 15-1 illustrates the effect of temperature on the speed of sound in water.

The pressure and salinity affect the speed of sound also but to a lesser degree than temperature. As the depth or pressure increases so does the sound velocity. An increase in salinity will also increase speed.

In echo ranging work, in which only the upper few hundred feet are involved, temperature is generally the most important factor causing variations in sound velocity. Salinity is relatively uniform in the open ocean and therefore of minor importance when dealing with sound velocity. Furthermore, in layers where vertical salinity gradients exist, there is nearly always a vertical temperature gradient.

In sound transmission the vertical velocity gradient is more important than the velocity itself, since it is the change in velocity with depth that determines how much refraction will take place. The velocity gradient is readily determined from the gradients of temperature and salinity. The salinity gradient is defined as the rate of change of salinity with depth in parts per thousand per meter. The temperature gradient is the rate of change of temperature with depth in degrees Celsius per meter. These gradients are therefore called positive if the quantity in question increases with depth, and negative if it decreases. In the majority of cases, temperature gradients in the sea are zero or negative. Moreover, except in certain localized areas, temperature gradients control the sound velocity gradients.

With a zero gradient in temperature (mixed layer) echo ranges are long because the sound rays are very nearly straight, having only a slight upward curvature due to the pressure effect. On the other hand, with a strong negative gradient

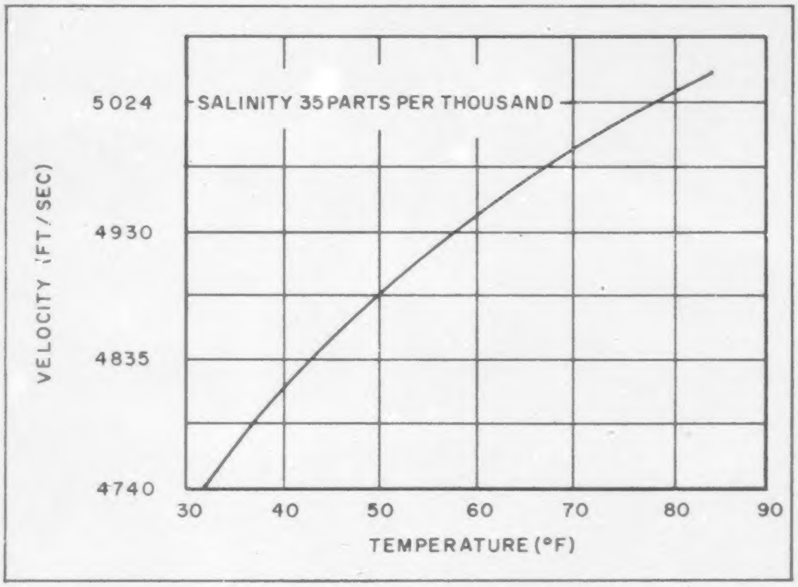


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Figure 15-1.—Effect of temperature on the speed of sound in water.

near the surface, echo ranges will be short because the sound beam is refracted sharply downward. In the ocean it is common to find a mixed layer (isovelocity) overlying a negative gradient. In such cases the echo range on an object in the mixed layer will be long, but the part of the sound beam that enters the negative gradient will be refracted downward, resulting in a reduction of range, and a split beam pattern.

With a strong negative gradient from the surface downward, each ray of sound beam curves down in a great arc, and that area beyond the horizontal limits of the beam is a so-called shadow zone into which no sound penetrates other than by scattering. An echo ranging vessel will not be able to detect an object in the shadow zone, but as soon as the object comes within the direct beam, the echoes will come in loud and clear.

With slighter negative gradients, generally with any gradient underlying a mixed layer, the shadow zone is not very clearly defined.

An object in the negative gradient beneath a mixed layer may be within the direct beam but still be undetectable because the echoes are too weak to be heard against the background or reverberation and ship's noise. This is known as the layer effect.

FORECASTING ADVECTION

The effect of the addition or removal of water by currents which result in the changing of the thermal structure of the water at a specific point is referred to as advection. In most cases this mass transport of the water masses will be accomplished by the ocean currents, of which there are varying types.

Wind-Driven Currents

The frictional drag of the wind sets up wind-driven currents which flow at less than 3 percent of the wind velocity. These wind-driven currents do not flow with the wind but are deflected 45 degrees to the right in the northern hemisphere and 45 degrees to the left in the southern hemisphere. This is caused by the earth's rotation and is closely related to its influence on the depth of mixing.

Permanent Currents

The redistribution of density resulting from the wind-driven current is what maintains the permanent current. Under the influence of the steady wind systems, such as the trade winds, in the lower latitudes and the westerlies in the higher latitudes, these permanent currents form the large scale current systems of the oceans. They are partly the indirect result of geographic differences in the heating and cooling of the water and partly the result of wind action. The character of the currents is also influenced by the configuration of the oceans, but in general there are clockwise gyrals in the northern hemisphere and counterclockwise gyrals in the southern hemisphere. Smaller currents exist near the continents. A countercurrent flows eastward between two westward flowing equatorial currents.

The permanent currents have several effects on the temperature conditions. Currents with poleward flow tend to carry warm water into cooler regions; conversely, equatorward flowing currents bring cooler waters into warmer regions. Within the currents themselves the distribution of density produces a temperature gradient such that, in the northern hemisphere, the water on the left side of a current has a lower average temperature than water on the right side. This may be reflected by a thinner mixed layer or even by lower surface temperatures. In the southern hemisphere the structure is reversed.

Divergence and Convergence of Surface Currents

Divergence of surface currents may occur under the influence of the wind. Examples of this are found along the western coasts of the continents and in the vicinity of the equator in the eastern parts of the Atlantic and Pacific. In these areas upwelling brings water toward the surface from moderate depths and the thermocline may be shallow or, in extreme cases, absent. The opposite effect, convergence occurs in the center of the subtropical gyrals in the northern and southern hemispheres. In these regions the surface water accumulates and consequently the thermocline may be very deep.

Tidal Currents

Tidal currents in partially isolated shallow areas have a marked effect on the temperature conditions because they also cause turbulent mixing. In the areas of strong tidal currents such as the English Channel, the water may remain virtually mixed throughout the year, although there is, of course, heating and cooling of the water column as a whole.

Internal Waves

Internal waves also affect the temperature distribution. The effect of these waves is reflected in a periodic rise and fall of the thermocline. Periods as long as 12 or 24 hours are known to exist and studies have shown that waves of only a few minutes' period may occur. Whether there is a continuous spectrum of frequencies is not known.

Methods of Forecasting Advection

Advection can be computed from knowledge of the temperature and ocean current field. Unfortunately the former is only moderately well defined in terms of the detail required for advection computations, and the velocity field is not only difficult to predict, but is not directly observed. There are two ways to regard advection: (1) by computing the change of temperature with time at the forecast point; or (2) by finding the source of the water expected at the forecast point and using the thermal structure of the source as the forecast.

The difference is illustrated in figure 15-2, where a forecast is desired for point A, and it is assumed that the 24 hour advection is from point B to point A. Recent BT observations are available for both points. One approach commonly used is to make all the heat budget and mixing modifications to the BT at point A, and as a last step the changes resulting from a different water structure being advected from point B are considered. This is not the most realistic approach as the water at point A will not be present in 24 hours. The water coming from point B represents the water that will be there instead.

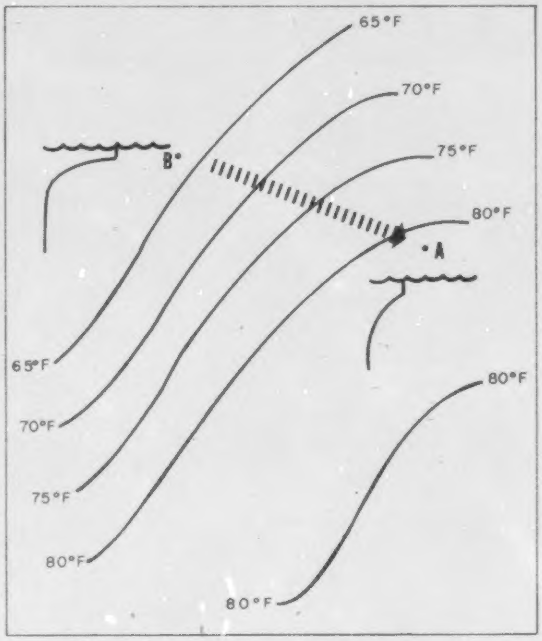

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Figure 15-2.—Typical advection.

The logical approach would be to assume the BT trace at point B represents the water that will be at the forecast point in 24 hours, and apply the heat budget and modifications to this trace instead. This is still not completely valid, since the entire water column from the surface to the bottom of the BT trace does not move at the same speed. The speed of the current generally decreases with depth, and in the case of wind drift, currents may be zero at the thermocline depth. This problem also arises in the first approach since advecting the whole thermal structure at the same speed would be in error.

The publication, Ocean Thermal Structure Forecasting, SP-105, Volume 5, outlines four conditions which cover most forecasting problems that will be encountered. Procedures and necessary graphs for predicting advection are contained within this publication. The procedures given are too lengthy to be contained within this manual; however, it is recommended that direct reference be made to the SP-105 publication for detailed information on forecasting advection.

FORECASTING THE HEAT BUDGET

The temperature structure of the ocean is determined primarily by its heat content, which is a constantly varying quantity. There is a continuous exchange of heat at the surface of the ocean. The ocean receives heat by absorption of the sun's radiation and by the condensation of water vapor in the air, when the water is colder than the air. The ocean loses heat by radiation to the atmosphere, by evaporation of water vapor when the water is warmer than the air, and possibly by conduction. Of the received heat, by far the largest quantity is due to the incoming solar radiation. Over the ocean as a whole it is balanced by the cooling resulting from reradiation and evaporation.

Incoming Radiation

The incoming radiation includes the invisible infrared and ultraviolet as well as visible light. Since it is received from the sun through the atmosphere, it obviously varies with latitude, season, time of day, and the atmospheric conditions, particularly the cloud cover. The total energy received during the year decreases with increasing latitude and in the lower latitudes of the tropical regions the seasonal variation is small, but with increasing latitude the difference between the amounts received during the summer and winter becomes very great. The effect of clouds is very pronounced; a heavy cloud cover may reduce the incoming radiation to less than 25 percent of that received on a clear day.

Direct heating of the water by the sun is limited to relatively shallow depths. Only about 3 percent of the radiation penetrates below 300 feet and over 50 percent (all the infrarred) is absorbed in the first few inches. If there were no compensating heat losses and no mixing fantastically high surface temperatures and extremely sharp negative gradients just below the surface would occur. The penetration of light varies somewhat from place to place depending upon the amount of suspended debris and organic pigments in the water. This applies to open ocean, since near shore and in areas of heavy plant growth the water is practically opaque to all wavelengths.

Effective Back Radiation

Effective back radiation is the term used for the excess of infrared emitted by the sea surface over that received from the air. This balances somewhat less than one-half of the incoming solar radiation, on the average. It decreases with increasing humidity and increasing cloud cover, and may increase or decrease with increasing water temperature. The latter is dependent on how much vaporization affects the water vapor content of the overlying air. With heavy, low lying clouds present, the effective back radiation drops to less than 25 percent of that on a clear day, largely because the clouds themselves are sources of infrared and radiate heat into the ocean on their own account. Clouds prevent direct solar radiation from reaching the sea surface. Heat losses from back radiation occur in the uppermost fraction of an inch in the water and are transmitted to greater depths by convective overturn and wind mixing.

Evaporation

Evaporation depends primarily upon the temperature of the water and the air, the humidity, and the wind strength. Evaporation can best be understood by considering the process as one of transfer of water vapor away from the surface. The greater the water vapor gradient the more rapid the evaporation and hence the greater the heat loss. Cold, dry air overlying warm water therefore favors rapid evaporation. High winds increase evaporation by removing the water vapor.

Forecasting Method

The heat budget will be represented by the following equation:

$$Q = Q_s + Q_c - Q_b - Q_e - Q_r - Q_h$$

where Q = Net gain or loss of heat
Q_s = Insolation
Q_c = Heat gain by condensation
Q_b = Effective back radiation
Q_e = Heat loss owing to evaporation
Q_r = Reflected radiation
Q_h = Heat conduction across interface

To determine the value of the individual components, it is necessary to compute values from graphs and nomograms. The publication, Ocean Thermal Structure Forecasting, SP-105, contains these values along with the necessary graphs and nomograms. It is recommended that forecasters utilize this publication when determining the gain or loss of heat to the ocean surfaces.

FORECASTING MIXING

To forecast the mixing that will occur in a portion of the ocean it is necessary to first discuss some of the mixing processes that take place.

Convective Overturn

When surface water cools, its density increases and it sinks, causing convective overturn. Equally important is the increase in salinity resulting from evaporation. The increased density, arising from this cause, contributes greatly to overturn and the development of isothermal surface layers. Thus, cooling by evaporation increases density in two ways and is less likely to be accompanied by positive temperature gradients than is cooling by radiation alone.

Conditions that tend to lessen the salinity of the surface layer would have the opposite effect, and would tend to favor the development of positive gradients. Such a condition might result from precipitation. For the ocean as a whole however, evaporation exceeds precipitation. This is shown in figure 15-3. It will be noted in the figure that regions of excess evaporation in low and mid-latitudes correspond to regions of relatively high surface salinity and deep thermoclines. Just north of the equator and in latitudes above 40 degrees, where precipitation exceeds evaporation, the surface salinity is low.

The deficit in the water content of the ocean that is caused by the general excess of evaporation over precipitation is made up by runoff from land. Near land, and especially near the mouths of rivers, surface salinities are lower than in the open oceans or at depths. This favors the development of positive temperature gradients, since it increases their stability.

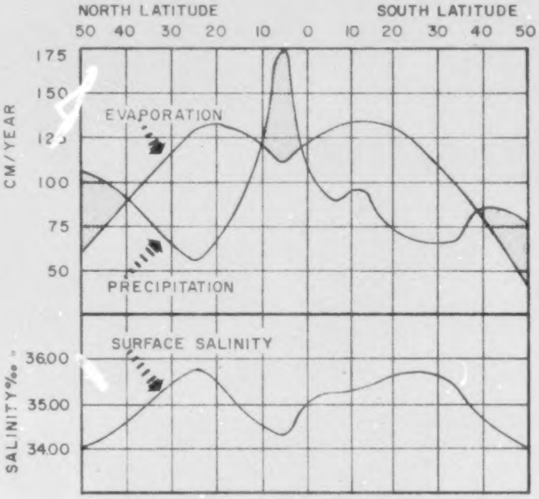

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Figure 15-3.—Variation of average evaporation, precipitation, and salinity with latitude. Shaded areas show regions where precipitation exceeds evaporation.

Mechanical Mixing

Mechanical mixing is caused by wind and does not necessarily involve any gain or loss of heat; nevertheless it may modify the temperature distribution. The effect of wind depends not only upon its strength, but also upon its duration and on the distance over which it has blown. It is quite obvious that the first effect of the wind will be confined to the immediate surface, but that the turbulence will extend to greater depths after the wind has been blowing for some time. The original density distribution of the surface layer will affect the rate at which the turbulence penetrates the layer. A very stable layer will be less easily mixed.

Rotation of the Earth

It is a remarkable fact that the daily rotation of the earth about its axis also affects the depth to which the wind mixing penetrates. The theories concerning the rotation effect are too involved to be included in this manual, however, all agree that a wind of given force will ultimately produce a deeper mixed layer in low latitudes than in high.

Forecasting Methods

Forecasting procedures for predicting mixing action over the ocean involves the computation of a number of variables. The procedures are discussed, with examples, in the publication, Ocean Thermal Structure Forecasting, SP-105. Personnel should refer to this publication for procedures and formulas for preparing such forecasts.

APPLICATION OF NUMERICAL PRODUCTS TO SUBSURFACE FORECASTS

In preparing forecasts for ASW operations forecasters will rely to a great extent on charts that are received via facsimile. Most of these will be numerically produced and they will include the Sea Surface Temperature Analysis (SST), the Sonic or Mixed Layer Depth Chart (SLD or MLD), the Gradient Below the Surface (GRD), Selected Bathythermograph Traces, Sea Height Prog Charts, and others whose availability may vary.

Sea Surface Temperature charts are constructed to provide a representative view of the sea surface temperature patterns. Only the main features of the SST field are shown. Diurnal and other short-term temperature changes are not normally detected. SST charts provide fundamental information for the conversion of environmental data into operational data of concern to ASW forces. This information may be readily utilized in forecasting of sonar ranges.

Sonic Layer or Mixed Layer Depth charts are based on BT reports received. MLD/SLD charts are used in conjunction with SST charts to determine the depth at which submarines may or may not be readily detected. Areas in which sonar ranging may also be affected can be determined.

The Gradient Below the Layer depicts the vertical temperature gradient in terms of change, plus or minus, that is taking place over a given distance, which in this case is depth. The chart delineates areas of change in degrees per 100 feet below the sonic layer. Temperature gradients provide a measure of the vertical sound velocity which may be utilized to determine the ray curvature and other variables.

Charts dealing with ocean surface conditions are prepared and disseminated also. Although sea condition is frequently ignored or lightly touched by forecasters, it is of great importance to the ASW operator, as it can reduce the effectiveness of equipment to a minimum. Combined sea height data is computed and disseminated in both facsimile chart and message form via the Fleet Broadcast. Data on sea state and sea waves are also prepared but not as widely disseminated. This information is available upon request.

A chart of selected BT traces is also received by some units. These are actual BT traces for selected points. Location may be given, keyed to operational orders, or the traces may be for predetermined points. The usage of the BT traces is somewhat limited unless they are in close proximity to the point for which the forecast is being prepared. They will, however, for their position, provide excellent data for estimating sonar conditions, by allowing the user to evaluate the temperature gradient in each layer of water.

Two of the most widely used numerical products used in the preparation of the ASW forecast are the ASRAPS and SHARPS which are received via message. Both of these products will be discussed in the next section of this chapter.

THE ANTISUBMARINE WARFARE ENVIRONMENTAL PREDICTION SERVICE

The Antisubmarine Warfare Environmental Prediction Service, commonly referred to as ASWEPS, was established during the early part of 1959 for the purpose of developing an integrated system of predicting and displaying parameters for antisubmarine warfare operations. It is designed to provide oceanographic analyses and forecasts utilizing up-to-the-minute data collected from various observation points.

Oceanographic data is collected by the Regional Oceanographic Net and the Mobile Oceanographic Net. The raw data is then processed into usable ASW oceanographic data and returned to the fleet users in the form of environmental forecasts, regional charts, and detailed area charts.

OCEANOGRAPHIC SERVICES

A variety of oceanographic services are available to fleet users. These include: (1) general oceanographic analysis and forecasts; (2) tailored products such as analysis and forecasts prepared for environmental effects upon specific ASW sensor systems; and (3) oceanographic outlooks for specific areas.

Fleet meteorological units and NWSED's modify and tailor data received from central computing points, then present them to the local users, primarily in the form of predicted sensor ranges.

The various charts, both analysis and prognostic, are utilized by the weather unit along with message type forecasts to prepare a standard ASW environmental briefing folder. This folder provides each flight crew with all the synoptic and forecasted oceanographic data as well as forecasted sensor ranges. It will usually be issued to the flight crews at the preflight weather briefing.

ASRAP and SHARPS

The Acoustic Sensor Range Prediction System (ASRAP) and Ship Helicopter Acoustic Range Prediction System (SHARPS) have been developed to provide a standard environmental and tactical acoustic range prediction system applicable to both active and passive systems. This range prediction information is normally received in message form, interpreted and presented to the user by Naval Weather Service personnel, except in the case of smaller ships where this is done by sonar operators.

ASRAP provides standard environmental and tactical predictions for patrol (VP) ASW forces for both active and passive systems. ASRAP displays are standard with the passive ranges being shown using a graph-like chart with decibels and range in nautical miles as the vertical and horizontal axis, respectively. Information is provided for both shallow and deep targets using hydrophones. The active ASRAP display presents forecasted ranges for active systems with varying combinations of Target/SUS and Hydrophone depths. The forecasts provided are for predetermined points of latitude and longitude These points are considered to be representative for the entire corresponding area surrounding them.

The SHARPS provides routine daily range calculations based upon propagation loss data for fleet sonar systems. The SHARPS system is area oriented, an advantage to both operational planners and tacticians. SHARPS provides a spectrum of ranges expected in an area. Bottom reverberation loss calculations, convergence zone, and bottom bounce ranges are provided for shipboard systems, while passive ranges for helicopter systems are also included. The system is flexible in that it is applicable to all active sonar systems in a variety of operating situations.

Both systems are under continuous evaluation and are subject to change to cover additional systems, take advantage of new data, or present a better display to the user.

CHAPTER 16

SPECIAL OBSERVATIONS AND FORECASTS

This chapter is intended to provide information related to the observation and forecasting of those environmental parameters which are difficult to categorize with the commonly recognized types of meteorological and/or oceanographic data.

AIR POLLUTION POTENTIAL

Air Pollution Potential (APP) is definable as a measure of the inability of the atmosphere to adequately dilute and disperse pollutants emitted into it based on values of specific meteorological parameters of the macroscale features. The Development Division of the National Meteorological Center (NMC) and the Division of Meteorology of the National Air Pollution Control Administration (NAPCA) have developed criteria and procedures to delineate areas on the macroscale in which high APP has the greatest possibility of occurring. As a result, air stagnation guidance data prepared by NMC is disseminated via facsimile and teletypewriter.

The Naval Weather Service Command is responsible for developing and maintaining up-to-date procedures related to interpreting and tailoring National Weather Service air pollution potential (APP) forecasts for interested U.S. Navy sea and shore activities. This section will present information relative to the use of APP forecasts and the procedures for tailoring them for local application.

The routine application of meteorology to local air pollution control is relatively new by comparison to other aspects of forecasting. It is incumbent upon the individual forecasters to apply their experience and judgement in the application of this data as part of their environmental support services. The National Weather Service has established Environmental Meteorological Support Units (EMSU) at major cities in the United States to assist in providing the necessary data required to maintain an efficient pollution forecasting program. Naval Weather Service personnel, as practicable, should establish a close working relationship with the EMSU Meteorologist.

The problem of air pollution is caused basically by a source emitting pollutants within a space which is unable to adequately dilute or disperse these pollutants. Both the pollutant source and the dispersion or dilutant capabilities of the area must be considered when determining air pollution potential. Some understanding of the nature of pollutants is therefore required.

POLLUTANTS

Pollutants are particles, gases, or liquid aerosols in the atmosphere which have undesirable effects on man or his surroundings. The magnitude of the concentrations of these pollutants ordinarily determines their undesirable characteristics. If, for example, air reaching an installation is pollution free, then the installation might be able to emit pollutants without exceeding undesirable concentrations even under restricted dispersion conditions. On the other hand, if the air reaching the installation is already saturated with pollutants, then even under good dispersion conditions, pollution emissions might not be advisable. Such cases can be resolved by predicting pollution concentrations through a comparison of existing diffusion properties of the atmosphere with the known operating characteristics of the plant, factory, ship, or other pollution source.

Dispersion Concepts

Atmospheric dispersion conditions are classified as good, moderate, or poor, the latter being a condition of high air pollution potential (HAPP).

The idea of the atmosphere dispersing, diluting, or ventilating pollutants is easily visualized. With an unstable lapse rate through a deep layer of the atmosphere and a strong wind, pollutants may be spread through an extensive volume of the atmosphere and diluted to minimal concentrations. On the other hand, a low inversion and a light wind may confine emissions to a shallow atmospheric layer and pollutant concentrations become larger. If the latter conditions persist and emissions continue, pollutant concentrations may become unacceptable.

APP Terminology

Discussing pollutant dispersion quantitatively requires the use of a number of descriptive terms with which the forecaster must become familiar. These terms are defined in the following paragraphs.

AIR POLLUTION POTENTIAL (APP).–A measure of the inability of the atmosphere to adequately dilute and disperse pollutants emitted into it, based on values of specific meteorological parameters of the macroscale features.

MIXING HEIGHT.–The surface-based layer in which relatively vigorous mixing occurs (meters).

TRANSPORT WIND SPEED.–A measure of the average rate of the horizontal transport of air within the mixing layer (meters per second).

VENTILATION.–The product of the mixing height and the transport wind speed. A measure of the volume rate of horizontal transport of air within the mixing layer, per unit distance, normal to the wind (meters^2 per second).

STAGNATION AREA.–A combination of stable stratification, weak horizontal wind speed components, and little, if any, significant precipitation. It is usually associated with a warm-core type anticyclone.

FAVORABLE CONDITIONS FOR POLLUTION

The three principal meteorological conditions which have been found to be most favorable for the formation of a pollutant stagnation area are as follows:

1. A slow moving anticyclone with a small horizontal pressure gradient.
2. Light surface winds not exceeding seven knots and winds aloft not exceeding 25 knots.
3. Subsidence in the lower layers of the atmosphere. This phenomenon with its attendant warning and drying effect produces stabilization and the formation of inversions which limit vertical mixing.

The greatest variations of air pollution potential are those of short duration due to the systematic variation of wind and stability between night and day.

As stated earlier in this chapter, atmospheric dispersion of pollutants is described as good, moderate, or poor in intensity. The intensity assigned depends on the mixing depth (MXDP) and the transport windspeed (TW). Whether these conditions will persist is governed by the presence or absence of a stagnation area. A synoptic situation with a deep unstable layer and a strong wind is described as having a large MXDP and a large TW; whereas, a synoptic situation with a low inversion and low wind speed is characterized as having a low MXDP and low TW. The product of MXDP and TW is termed ventilation and is a measure of the atmosphere's capability to dilute or to disperse pollutants. The use of the term "stagnation area" provides an objective delineation of a geographical area where the atmosphere will undergo little synoptic change and where pollutants will accumulate. It is an important forecast parameter and will be discussed in more detail later in this chapter.

Parameters and Critical Values for Delineating Stagnation Areas

Computations for delineation of stagnation areas are accomplished at NMC primarily by computers. Furthermore, as a result of the relative newness of APP forecasting, parameters and critical values are continually being reevaluated and are subject to change. In view of these facts, it is not feasible to list specific items in this publication. It is sufficient to state that wind speed, stability criteria, and precipitation are major considerations.

Calculation of Mixing Height and Transport Windspeed

Once the stagnation areas have been determined, the next step is to calculate the mixing height and the transport wind speed. These computations as well as those previously mentioned for stagnation areas are normally accomplished objectively by computers at NMC.

NMC AIR POLLUTION POTENTIAL PRODUCTS

National Meteorological Center APP products transmitted over facsimile and/or teletype circuits and the dispersion criteria derived from them identify meteorological conditions associated with the large-scale buildup and dispersion of pollutants over or downwind of an urban area. Such areas emit pollutants from ground-level sources as well as from elevated sources, such as stacks, etc. Downwind of the sources and over the areas where pollutants from all sources mix together, pollution concentrations correspond well with observed APP conditions. Keep in mind however that current and forecast meteorological conditions must be considered when utilizing these products. For example, if a very stable ground inversion exists below stack height outside of an urban area, it is very probable that stack emissions would drift over the area above the inversion with little effect on ground pollution concentrations. Pollution concentrations would probably remain low and the area might be considered pollution-free despite an APP forecast calling for poor dispersion conditions. On the other hand, in an urban area where ground-level sources of pollution from high volume vehicular traffic, numerous home and commercial heating installations, etc., exist, these factors would make up for the reduction in pollution from the elevated sources and the APP forecast would verify. Forecasts of APP are, thus, most useful in describing dispersion condition areas within or downwind of urban areas. However, by tailoring the forecasts in a manner similar to that described later in this section they may be applied to other locations.

In the preceding paragraphs we have discussed dispersion of pollutants in general terms. In the following paragraphs information is presented relating to obtaining Air Pollution Potential products over the facsimile and teletype circuits and the application of this data to forecasting.

Facsimile

Facsimile charts depicting APP information are transmitted daily over the NMC FOFAX circuits at various times depending on the circuit. A detailed description of these charts may be found in Technical Procedures Bulletin No. 69 (or its revision) available from U.S. Department of Commerce, NOAA, NWS, Silver Spring, Maryland 20910.

The data entered on the four panel facsimile chart will be similar to that illustrated in figure 16-1. Again, the newness of the program makes the data subject to change. The data will primarily be used to assist in making HAPP advisories. To a limited extent, the data may also be used in making local forecasts of good and moderate dispersion conditions.

APPLICATION.–The APP charts have two applications. Primarily, it alerts the forecaster to poor dispersion conditions and to the need for a HAPP advisory. Secondarily, it can be used to a limited extent as a data source for issuing advisories of good and moderate dispersion. To use the chart as a basis of a local HAPP advisory, the forecaster examines the chart to determine if his ship or station will be in an area conducive to the accumulation of pollutants. To check local TW or MXDP values, the forecaster can consult the detailed mesoscale information of FKUS-1, the local Environmental Meteorological Support Unit (EMSU) sounding, or a nearby radiosonde. Assuming HAPP criteria are met, an advisory is issued to the appropriate control authorities.

All Naval Weather Service units receive via Comet III circuitry the FKUS 1-Air Pollution Potential Data and FKUS-2 Air Stagnation Narrative. Detailed information on FKUS 1 is

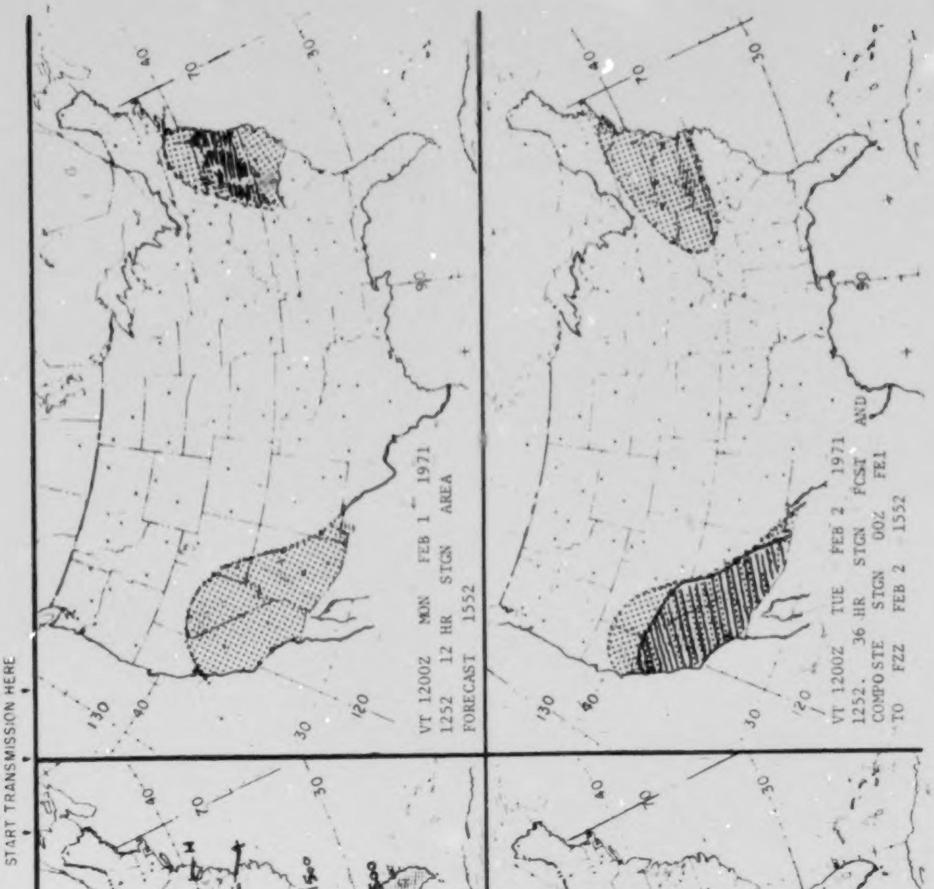
START TRANSMISSION HERE
0000Z MON FEB 1 1971
1252. OBSVD STGN AREA
TEST CALO MIXING HOT AND
TRANSPORT WIND SPEED 1552
<4MPS
4MPS
<1500M
>1500M
1500
VT 1200Z MON FEB 1 1971
1252 12 HR STGN AREA
FORECAST 1552
VT 0000Z TUE FEB 2 1971
1252. 24 HR STGN AREA
FORECAST 1552
VT 1200Z TUE FEB 2 1971
1252. 36 HR STGN FCST AND
COMPOSITE STGN 00Z FE1
TO FZZ FEB 2 1552
130
120
90
70
40
30

Figure 16-1.—Example of four panel APP facsimile chart.

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found on page 22 of ESSA Technical Memorandum WBTM 47, and on FKUS 2 in Technical Procedures Bulletin No. 58 (or its revision).

FKUS 1. This coded message provides detailed MXDP and TW information for 00Z, 12Z, and 00Z; i.e., data from yesterday afternoon, this morning, and this afternoon. Note that this afternoon's 00Z data are really a 12-hr forecast based on the 12Z sounding and winds, and that a 999 group indicates missing data.

FKUS 2. This plain language narrative message describes the facsimile package.

APPLICATION OF MESSAGE. These two messages allow calculation of all dispersion conditions and permit issuance of a HAPP advisory. Message FKUS 1 contains MXDP and TW data, and ventilation rates can be calculated using these inputs. Forecasting beyond 12 hours is aided by noting the stagnation areas and using a persistence forecast for TW and MXDP in these areas. The inclusion of yesterday's data also allows a check on the persistence forecast; the forecaster can estimate 24-hour changes and judge the possibility of persistence holding for another 24 hours.

CONVENTIONAL FACSIMILE PRODUCTS

In the absence of NMC APP products; conventional facsimile charts are of value in estimating and forecasting pollution conditions. Lower level winds can be estimated from winds aloft charts and many of the criteria for delineating a stagnation area can be determined from the vorticity and constant pressure charts. Available trajectory forecasts can also help in determining stagnation areas in forecasting lapse rates.

TAILORING THE FORECAST

The NMC APP products can be tailored for local application. Some factors to be considered are:

1. Emission Source. APP criteria are most accurate in predicting dispersion conditions over and downwind of a large urban pollution source. The closer the atmospheric characteristics of the ship or station resembles an urban area, the more likely is the APP forecast to be accurate over all dispersion conditions. APP forecasts of very good or very poor conditions, however, are likely to verify at most Naval installations.

2. Meteorological Conditions. TT and FAX data are usually calculated for points distant from the base. These data can be adjusted for local effects such as exposed locations with higher than prevailing windspeed elevation, etc. The local forecaster can usually improve the representativeness of MXDP and TW when local data are available or when local predictions are more representative. Such modifications must be used to make observed or forecast APP more representative of the conditions surrounding the ship or station.

3. Topography. Locations adjacent to mountains or ridges, for example, may find air movement blocked and that pollutants accumulate despite otherwise excellent dispersion conditions. Forecast APP conditions should be adjusted accordingly, with perhaps a different weighting factor used for a TW from a particular direction.

4. Synoptic Situation. Local interested activities may be alerted to anticipated changes in the synoptic situation to enable them to prepare for improved dispersion conditions. They may also be warned in advance of impending HAPP conditions.

5. Wind Direction. The local APP index may be adjusted for wind direction. If, for example, base pollution sources are to the west of the base, then perhaps the APP index during an east wind should indicate a lower APP index, i.e., better dispersion conditions, than during a west wind.

6. Timing. The forecaster should tend to be conservative in issuing a forecast of HAPP and be equally as conservative in calling for its end.

Figure 16-2 illustrates the possible format for an air pollution forecast worksheet. This is included as merely a suggested format and should be altered as necessary to meet local requirements.

It must be emphasized that forecasts are prepared for pollution potential and not for specific pollutant concentrations. Procedures for disseminating pollution forecasts should be developed in coordination with base pollution control authorities.

SUGGESTED AIR POLLUTION FORECAST WORKSHEET

1. Obtain appropriate FOFAX chart, teletype messages FKUS-1 and FKUS-2, and local EMSU and radiosonde reports for the day.

2. Determine weighting factor (WF) for morning conditions:

   a. TW = ________ & TW (WF) = ________

   b. MXDP = ________ & MXDP (WF) = ________

3. Determine weighting factor for afternoon conditions.

   a. TW = ________ & TW (WF) = ________

   b. MXDP = ________ & MXDP (WF) = ________

   c. Vent = MXDPXTW = ________ & Vent (WF) = ________

4. Calculate morning APP Index.

   TW (WF) + MXDP (WF) = ________

5. Calculate afternoon APP Index.

   TW (WF) + Vent (WF) = ________

6. If APP Index is between:

   a. +1 & -1, then forecast moderate dispersion.

   b. -1 & -4, then forecast good dispersion conditions.

7. Determine if base to be in a stagnation area for 36 hours by examining FOFAX chart or FTUS. If the base is so affected, and morning and afternoon APP index > +1, then forecast HAPP.

8. Determine end of HAPP conditions by noting when ventilation exceeds 8000 m^2/sec or TW exceeds 4.5 mps.

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Figure 16-2.—Suggested air pollution forecast worksheet.

HIGH AIR POLLUTION POTENTIAL (HAPP) FORECASTS

A HAPP advisory is the most important of the pollution forecasts. Conceivably, it could affect aircraft or ships operations and, certainly, it is the one to generate the widest interest with the public. The latter is especially true if a nearby city is under a pollution alert as a result of a HAPP forecast.

One system in use for determining if an advisory should be issued relates an APP index to dispersion as follows:

APP Index	Dispersion
+3	Poor, i.e., HAPP
+2	
+1	Moderate to Poor
0	Moderate
-1	Moderate to Good
-2	Good
-3	
-4	

For measurable precipitation, subtract 1 from the APP index if the morning or afternoon APP is greater than zero.

The APP index is determined by using MXDP and TW to determine weighting factors which are then added separately for morning and afternoon conditions to obtain a measure of air pollution potential. The weighting factor criteria are provided in Table 16-1. The morning APP index is equal to the Weighting Factor for TW plus the Weighting Factor for MXDP. The afternoon APP index is equal to the Weighting Factor for TW plus the Weighting Factor for Ventilation (Vent); Vent is determined by multiplying MXDP by TW.

The dispersion criteria provided in table 16-1 are included for use as a guide only and are for locations away from an urban area. These criteria normally require modification to meet individual requirements. In the absence of urban pollution, dispersion conditions defined as good by these criteria should verify while those defined as bad may not be quite as bad as a city area under similar meteorological conditions.

Table 16-1.–Pollution dispersion criteria.

Minimum dispersion period (morning)	
Weighting Factor for Mixing Depth (MXDP)	Weighting Factor for Transport Wind (TW)
$\text{MXDP} < 250\ \text{m} = +1$	$\text{TW} < 2\ \text{mps} = +2$
$250\ \text{m} \leqslant \text{MXDP} < 500\ \text{m} = 0$	$2\ \text{mps} \leqslant \text{TW} < 4\ \text{mps} = +1$
$500\ \text{m} \leqslant \text{MXDP} \leqslant 700\ \text{m} = -1$	$4\ \text{mps} \leqslant \text{TW} < 6\ \text{mps} = 0$
$\text{MXDP} > 700\ \text{m} = -2$	$6\ \text{mps} \leqslant \text{TW} \leqslant 8\ \text{mps} = -1$
	$\text{TW} > 8\ \text{mps} = -2$
Maximum dispersion period (normally afternoon)	
Weighting Factor for Ventilation (Vent)	Weighting Factor for Transport Wind (TW)
$\text{Vent} < 4000\ \frac{\text{m}^2}{\text{sec}} = +1$	$\text{TW} \leqslant 2.5\ \text{mps} = +2$
$4000\ \frac{\text{m}^2}{\text{sec}} \leqslant \text{Vent} < 6000\ \frac{\text{m}^2}{\text{sec}} = +0$	$2.5\ \text{mps} < \text{TW} \leqslant 4.0\ \text{mps} = +1$
$6000\ \frac{\text{m}^2}{\text{sec}} \leqslant \text{Vent} < 8000\ \frac{\text{m}^2}{\text{sec}} = -1$	$4.0\ \text{mps} < \text{TW} \leqslant 5.0\ \text{mps} = 0$
$8000\ \frac{\text{m}^2}{\text{sec}} \leqslant \text{Vent} = -2$	$5.0\ \text{mps} < \text{TW} \leqslant 6.0\ \text{mps} = -1$
	$\text{TW} > 6\ \text{mps} = -2$

Criteria for Issuing a HAPP Advisory

Suggested criteria for issuing a local area HAPP advisory are as follows:

1. Detachment information indicates the following:
   - a. Local HAPP currently exists.
   - b. It is expected to persist for at least an additional 36 hours.
   - c. Objective threshold HAPP criteria as follows:
     - (1) Morning APP will be greater than +1.
     - (2) Afternoon APP will be greater than +1.
     - (3) Stagnation is forecast for 36 or more hours.
   - d. The ship or station is adjacent to or in close proximity to a city under a HAPP advisory and meteorological conditions at the ship or station are similar to those of the nearby city.

Typical Use of HAPP Advisory by Control Agency

The use of a HAPP advisory in a pollution alert system is illustrated in figure 16-3.

Termination of HAPP Advisory

A HAPP advisory will normally be terminated whenever ventilation exceeds 8,000 m^2/sec or the transport wind exceeds 4.5 mps.

RADAR INTERPRETATION

The application of radar as an aid to observing and forecasting weather has provided the forecaster with information of inestimable value in

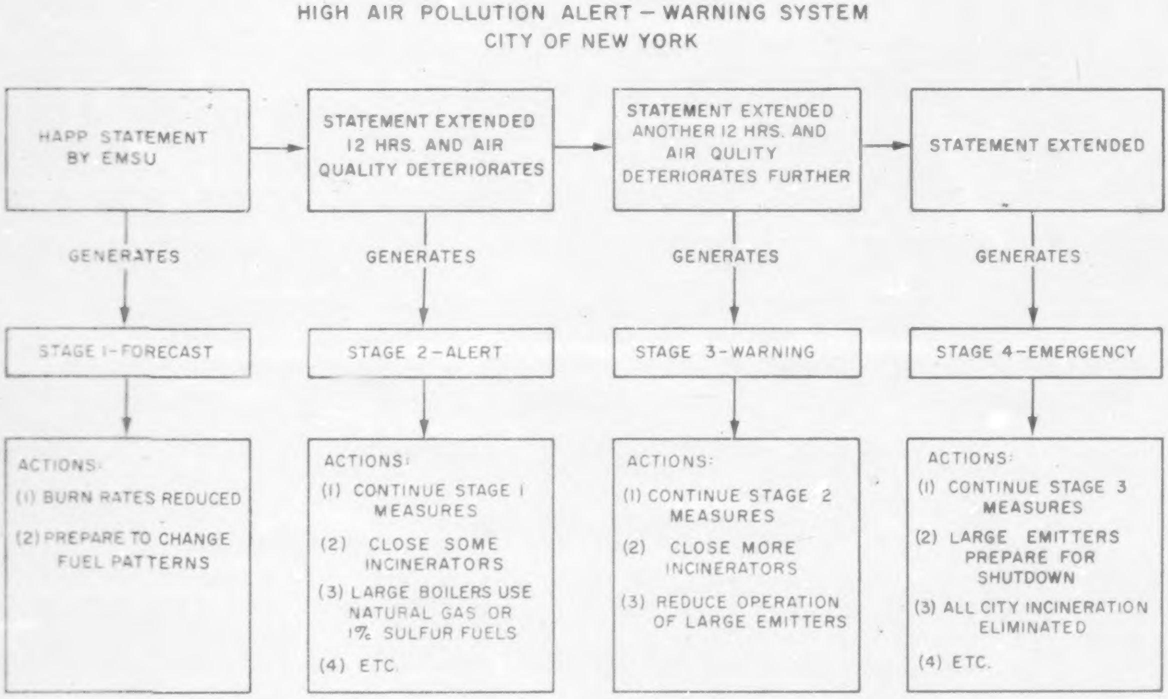
HIGH AIR POLLUTION ALERT — WARNING SYSTEM
CITY OF NEW YORK
HAPP STATEMENT BY EMSU
STATEMENT EXTENDED 12 HRS. AND AIR QUALITY DETERIORATES
STATEMENT EXTENDED ANOTHER 12 HRS. AND AIR QULITY DETERIORATES FURTHER
STATEMENT EXTENDED
GENERATES
GENERATES
GENERATES
GENERATES
STAGE 1–FORECAST
STAGE 2–ALERT
STAGE 3–WARNING
STAGE 4–EMERGENCY
ACTIONS: (1) BURN RATES REDUCED (2) PREPARE TO CHANGE FUEL PATTERNS
ACTIONS: (1) CONTINUE STAGE 1 MEASURES (2) CLOSE SOME INCINERATORS (3) LARGE BOILERS USE NATURAL GAS OR 1% SULFUR FUELS (4) ETC.
ACTIONS: (1) CONTINUE STAGE 2 MEASURES (2) CLOSE MORE INCINERATORS (3) REDUCE OPERATION OF LARGE EMITTERS
ACTIONS: (1) CONTINUE STAGE 3 MEASURES (2) LARGE EMITTERS PREPARE FOR SHUTDOWN (3) ALL CITY INCINERATION ELIMINATED (4) ETC.

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Figure 16-3.—Example of utilization of HAPP advisory by a controlling agency.

many instances. Its use in thunderstorm, tornado, and hurricane detection and warning have materially reduced the destruction and loss of life due to these phenomena. Meteorological radar is discussed in chapter 18 of this manual. Other types of shipboard radar, although not specifically designed to observe weather, can be useful to the forecaster during adverse weather if its current operation is not of a higher priority. The value of this equipment, as with all other meteorological equipment, is to a large degree dependent upon the experience of the operators and forecasters involved. The intent of this section is to present basic information related to the interpretation of radar echoes, thereby providing a basic foundation upon which the forecaster may build with increasing knowledge and experience.

IDENTIFICATION OF WEATHER ECHOES USING A PPI-SCOPE

Although a synoptic map gives definite indication of an approaching front or hurricane, or of the presence of thunderstorms in a specific area, minute-to-minute tracking of these weather phenomena is not possible. From the usual source of weather information, the exact time of advent of adverse weather cannot be forecast, although under favorable conditions it can be approximated. The reflection of radar pulses from clouds associated with precipitation permits the continuous tracking of the position of such clouds with respect to the location of the station. Thus, a degree of accuracy in the forecasting of the approach of unfavorable weather, not possible by standard methods, can be achieved using radar methods.

It cannot be overemphasized that all available scopes should be used in studying weather phenomena. Refer to chapter 18 of this manual for a description of the various meteorological radar scopes. This section is written, however, as though the PPI were the only available scope.

The Plan Position Indicator (PPI) scope should be used to determine the character of the echoes which are classified as isolated, widely scattered area, scattered area broken area, solid area, line of widely scattered echoes, line of scattered echoes, broken line of echoes, solid line of echoes, spiral band area, stratified elevated echo, and fine line. Table 5-1 in Weather Radar Manual, NavAir 50-1P-2, is used to determine the characteristic of an echo area. Figure 16-4 illustrates the appearance of a solid line of echoes on a PPI scope.

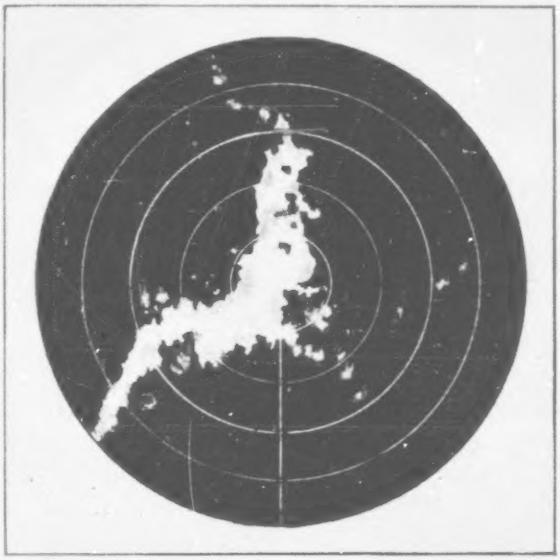

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Figure 16-4.—Line Echo Wave Pattern (LEWP) in a squall line as it appears on a PPI scope.

Scattered echoes may be related to air mass weather; lines of echoes may be related to squall lines or cold fronts; certain characteristic curves in echo lines may be related to frontal waves; and widespread relatively homogeneous echo patterns may be related to warm frontal precipitation under stable conditions (banded structures under less stable conditions).

Thunderstorms

Consider, for the moment, the typical physical appearance and behavior of a thunderstorm. Usually, the onset of rain is quite sharp and the precipitation is heavy. As quickly as a thunderstorm passes, the rain stops. It is not surprising, therefore, that the echo which is returned from a thunderstorm is almost always bright (good raindrop targets) and reasonably sharp edged (very little light rain, which gives marginal echoes). The brightness and the sharpness of a thunderstorm echo distinguish it from almost any other type of echo.

The development of the thunderstorm can be followed by the Aerographer's Mate quite clearly on the PPI. The horizontal extent can be determined by watching how many degrees of azimuth are covered by the echo. If there is only one thunderstorm in the area, sector scan is desirable, since the time taken by the antenna in rotating 360° need not be lost.

The location of the active cells within the thunderstorm can be found by using the gain control. The higher the gain setting, the higher is the power of the set. The highest setting shows up everything that the radar is capable of seeing. As the gain is reduced, the less intense parts of the echo drop out. Since the cells are the most active part of the thunderstorm, they are the last to disappear as the gain is reduced.

Rain or Snow

Since water droplets scatter about five times as much energy as corresponding snow crystals, the return from snow tends to be weaker, and the differences of intensity within a snowstorm are generally much less than in a rainstorm. Both texture and behavior help in distinguishing rain echoes from those of snow. A typical PPI presentation of snow is a uniform hazy or coarse echo with very diffused edges. The texture of snow echoes is often described as soft in contrast to the sharp or hard echoes produced by rain.

Another clue as to what kind of echo is involved may be found by varying the gain setting. There seems to be much more variation in the intensity of rain than of snow. A rain echo, as the gain is reduced, does not disappear as a unit. Certain portions fade out much more quickly than others. With snow, there is much greater likelihood that the entire echo will disappear at once. This characteristic may be used to good advantage in identifying the nature of the echo-producing substance.

Cold Fronts and Squall Lines

In considering the appearance of cold fronts and squall lines on a PPI-scope, start with a concept of how these phenomena appear in nature. A cold front with nothing in it contains no precipitation particles and does not show up on radar. A well-defined active cold front and particularly a squall line contain convective clouds and are marked by a band of active precipitation. On the PPI, all areas of precipitation will show up. The typical, unmistakable appearance of a front or squall line is that of a narrow band of discrete echoes oriented in a line, moving across the scope as a unit. (See figure 16-4.)

Experience seems to indicate that this band of echoes is closely associated with the frontal position as drawn on a weather map. It also indicates, however, that the frontal position and the cloud position do not necessarily coincide. The front on the map, drawn by skilled analysts, may be in advance of or behind the cloud band. Sometimes, the two lines coincide. During the passage of a front across the radar station, when tracking is effected from one side of the scope to the other, it is apparent that even for the single front, the position of the clouds relative to the front varies with time.

Warm Fronts

The classic picture of a warm frontal region is that of a wide cloud shield containing layered, precipitating clouds with perhaps a few thunderstorms penetrating the layers. On the PPI, the warm frontal picture is substantially the same. A large part of the scope may be covered by soft echoes (indicating continuous rather than showery precipitation). Usually, at full gain setting, thunderstorms even if present do not show up. They are hidden in the continuum of the warm-front echoes. With reduced gain setting, the overall echo tends to disappear and only the thunderstorm cells remain.

Hurricanes and Typhoons

Each time another tropical storm is tracked by radar, a new characteristic of such storms is observed. Any attempt to give a valid description of the appearance of a hurricane or typhoon on radar would undoubtedly be in error. On radar, as in every other way, each tropical storm seems to be an individual. Nonetheless, a generalized description can be given with the understanding that any specific storm may be an oddity.

Assume that a hurricane is approaching the radar set, passes close to the station, and then moves off. Until the rain shield associated with the storm is within approximately 250 miles of the radar, regardless of the power of the set, the PPI will not show any evidence of the storm. This is the result of the fact that the earth's surface is curved so that even when a radar beam leaves the set in a horizontal path, it exceeds the height of the hurricane clouds beyond about 250 miles. As the storm continues to approach, echoes begin to appear on the scope. These look almost identical with those produced by a warm front and, if knowledge of the presence of a hurricane were not available from other sources, they would be mistaken for warm frontal clouds. As the hurricane continues to come closer, the echoes begin to develop in distinctive patterns. They acquire structure and appear as a series of concentric bands. (See figure 16-5).

The position of the eye of the storm may be approximated by finding the center of curvature of the bands. Actually, the bands seems to spiral about the eye; and as more and more of the hurricane is seen on the scope, the spiral pattern becomes more evident. The eye of the storm is a blank area on the scope. No precipitation occurs within the eye. The relationship between the close-in spiral bands and the eye is not clear. It appears that the eye shifts, forms and re-forms, or develops asymmetrically with respect to the nearest precipitation bands. As the storm recedes, the trailing half of a tropical storm picture is far less clear, since the trailing half of a tropical storm usually contains less precipitation than the leading half.

Lightning

It has been observed that lightning strokes show up on the PPI. Usually, the strokes cannot be identified by the naked eye. Their persistence is too short for spotting and identification. Motion pictures of the PPI with active thunderstorms do show the lightning strokes. When seen, there can be no mistake in their identification. On the scope, they resemble exactly their appearance in the sky.

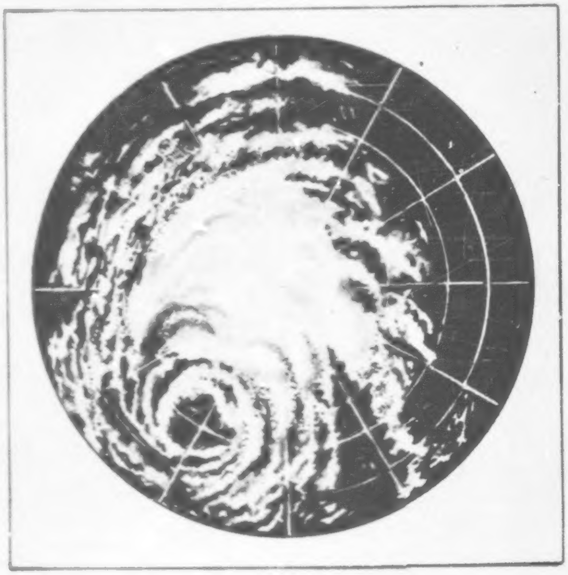

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Figure 16-5.—Hurricane showing echo-free eye on PPI scope.

Tornadoes

One of the most controversial issues in radar meteorology is the identification of tornadoes on a PPI. In general, the contention is that the tight circulation associated with a tornado shows up as a hook, V, or eye on an intense thunderstorm echo. There are many instances in which a PPI picture was taken of such a manifestation while a tornado was in progress. The location of the hook or V coincided exactly with the known location of the tornado. Unfortunately, hooks, V's and eyes are also seen on intense thunderstorms when tornadoes are not present. In a still picture, those associated with the destructive storms are no different from those which are not. Only when the manner of hook development is watched over a period of time can an experienced observer distinguish between tornado hooks and meaningless hooks—and even then, there may be more or less doubt.

PLOTTING MOVEMENT OF WEATHER ECHOES

Weather echoes may be plotted in several ways. A suggested method is to use an acetate overlay. To do this, first, cut to size an acetate sheet that will lie flat over the PPI-scope. Next, mark the four directional reference points (N, E, S, and W), and then a center reference point to lay over the center of the scope.

When severe weather is approaching the station, lay the acetate overlay over the scope and at designated time intervals, depending on the speed of movement of the weather, place reference marks along the weather echoes. The movements can then be extrapolated to determine which storms, if any, are likely to affect the terminal. Obviously, attention should be focused on potentially critical upstream areas. When extrapolating the movement over the terminal, allowance should be made for modification by significant local terrain effects and for tendencies of the area to change in size and intensity as indicated by successive radar observations.

IDENTIFICATION OF WEATHER ECHOES USING AN RHI-SCOPE

The scope which shows range height indication can provide very valuable meteorological information which is not available in any other way. The presentation of the RHI is a vertical cross section of the atmosphere. When the RHI is used, the antenna is fixed with respect to azimuth and permitted to scan in elevation only.

Bright Band

Observers using an RHI-scope frequently speak of the bright band which appears on their scopes and relate the phenomenon to the 0° isotherm. The proper procedure for observing this phenomenon is for the radar operator to slowly reduce the receiver gain while the antenna is scanning in elevation. Also, short range will give the best results in this type of observation.

A bright band is the result of the simultaneous existence of snow, melting snow, and rain in a vertical cross section of the atmosphere. At upper levels, the precipitating particles are snow crystals. Snow is a relatively poor reflector of a radar beam. In the higher levels of the atmosphere, therefore, where the snow predominates, the echo return is poor. Rain droplets, on the other hand, make relatively good reflectors and the signal return is high. A well-defined and thin horizontal bright band is indicative of very stable air. Under extremely unstable conditions, however, the layer represented by the bright band becomes so deep and mixed up that there is little or no effect noticeable on the radarscope.

The Aerographer's Mate should note from the discussion above that the bright band is not the position of the 0° isotherm. When the relative rate of fall of snow and rain is superimposed on the differential reflectivity of the two, the bright band may be explained. Snow falls at a slow rate (about 0.5 mps); rain falls at a fast rate (about 4 to 8 mps). During the time in which a snow particle melts to form a rain droplet, it has the approximate rate of fall of snow, but the reflectivity of rain. Due to the slow rate of fall, there is high concentration of particles of poor reflectivity in the snow region. This results in a relatively small signal return.

While the particles fall through the 0° isotherm and melting starts, there is a high concentration of particles for a limited zone (as in the snow) with good reflectivity (as in rain). This means a high signal return.

Below the melting layer, there is a lower concentration of particles due to the high rate of fall of the rain. Even though the rain has high reflectivity, the lower concentration of rain results in a reduced signal return. The region of highest signal return corresponds to the melting layer. It is the bright band.

The melting process which produces the bright band only starts at the 0° isotherm and the bright band maximum must be below the 0° level. The usual position of the bright band is approximately 1,500 feet below.

Atmospheric Stability

Careful examination of the bright band may yield some information about the stability of that portion of the atmosphere around the melting level. With very stable air, the band is clearly defined. It is narrow vertically but of wide horizontal extent. As the atmosphere becomes less stable, the distribution of precipitation becomes cellular and the band broadens vertically. With strong convection extending through the 0° level, the melting and freezing processes become mixed through so deep a layer that nothing resembling a bright band can be detected.

Cloud Types

In general, it is rather simple to identify the various cloud types which produce echoes on the RHI-scope. Instead of discussing these types with descriptive terms, it is probably desirable to reproduce RHI pictures of some of the clouds which may be seen on RHI. A number of these echo patterns are illustrated in figure 16-6.

In figure 16-6 (A), for example, the broken pattern shows convective type showers. The fact that the echo does not extend above 11,000 feet indicates that the showers are weak.

Figure 16-6 (B) shows shower cells in the lower levels. These showers seem to originate in a great cloud layer extending from about 12,000 to 26,000 feet.

Figure 16-6 (C) shows a definite series of stratified layers between 8,000 and 18,000 feet. The relative intensity of the layers could be determined by decreasing the gain progressively. Only the main layer would remain at the lower gain settings.

A thick layer directly above the station is shown by the RHI in figure 16-6 (D). A short distance away, the precipitation reaches the ground. The advance of the precipitation with time can be followed by watching the blank area near the station. The streaky nature of the echo in the upper levels indicates shower activity.

In figure 16-6 (E), three layers are pictured with precipitation falling through them. Again, a cellular structure seems indicated.

Thunderstorms and Turbulence

Since the RHI gives considerable information concerning thunderstorms, and since a great deal of knowledge is available relating thunderstorms and turbulence, it is apparent that the RHI portraying a thunderstorm gives a fairly complete picture of associated turbulence (and, incidentally, icing and damaging hail).

In an actively developing thunderstorm, turbulence is most severe. During the time that the top of the radar echo rises rapidly, the thunderstorm is most active. The higher the top of the storm, the more violent is the storm and the greater is the attendant turbulence.

A decaying storm can be seen as decreasing in altitude with time. This subsidence of the top of the storm shows that convective activity is decreasing and turbulence is decreasing as well. The bright band develops as turbulence decreases.

Upper-Level Winds

A careful study of the RHI will frequently provide valuable information concerning winds aloft. When precipitation shows any shower characteristics, the cells can be observed in the RHI scope as separated columns of falling rain or snow. These columns often slant out of the vertical because of wind shear.

Examine figure 16-7 closely. This is a typical RHI picture from which some indication of wind structure can be obtained. Note that the picture shows a curving streak which begins at high altitude and extends downward.

Consider a cloud moving with the wind at cloud level. (See fig. 16-8.) As the cloud moves along, the rain or snow particles drop from it and fall progressively to lower levels. Track these individual particles. Label each particle with a letter starting with "a" and each time interval with an Arabic numeral starting with time "1." Thus, in the notation c.4, the dot between the "c" and the "4" represents the particle, while the letter "c" shows that it is the third particle to be dropped. The "4" indicates that four units of time have elapsed since the start of the observation. Label with Roman numerals the layers of the atmosphere through which the particles drop.

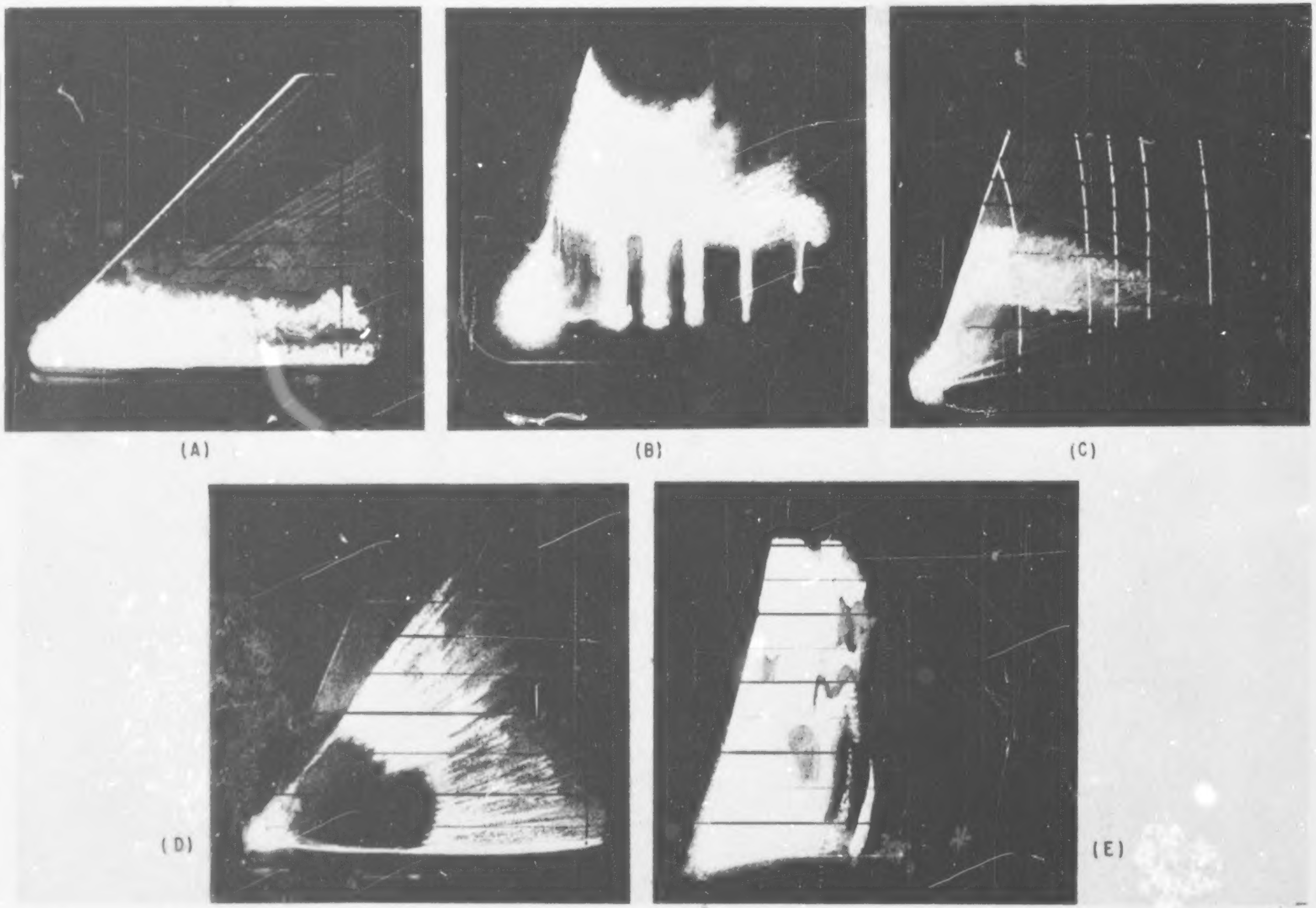

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Figure 16-6.—RHI scope precipitation patterns.

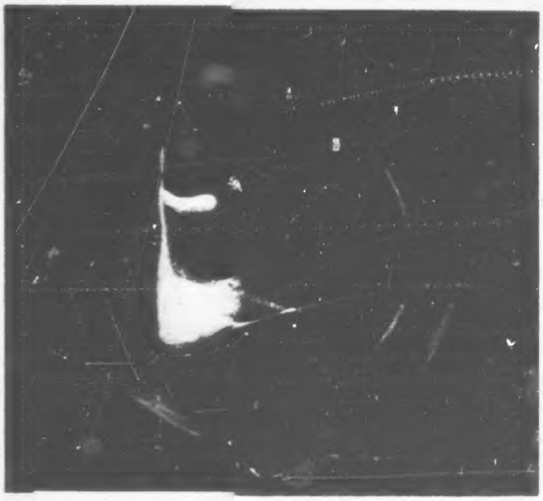

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Figure 16-7.—RHI picture with some indication of wind structure.

Figure 16-8 shows the generation of precipitation patterns in a wind field. The wind field assumed is indicated at the right side of the figure by arrows. The wind direction is in the plane of the paper (page) going from right to left. The length of the arrow indicates the wind speed. The double arrow at the top shows the speed of the generating cloud.

The dashed lines show the trajectories of some of the individual particles. Consider particle "a." It falls from the cloud at time "1." As it falls through atmospheric level VIII, it is under the influence of the mean wind in that layer. It is displaced horizontally to the left so that it reaches the bottom of layer VIII at time "2." The fall through layer VII results in continued displacement toward the left (although a smaller one) since the wind in layer VII is less than that in VIII. It is now displaced to the position it occupies at time "3." It will continue to be displaced dependent on the wind speeds the particle encounters on its descent.

Particle "b" is generated later than particle "a." Let us say that it is released from the cloud at time "2." Thus, it starts its fall through layer VIII just as particle "a" leaves that layer. The trajectory of particle "b" is identical with that of particle "a" except that the particle is delayed. Particle "c" is further delayed, as is each new particle falling from this cell.

RADAR WAVE PROPAGATION AND REFRACTIVE INDEX

Propagation means the travel of electromagnetic waves, as sent from the radar antenna, through a medium. Dissemination of radar waves is another way of expressing it.

Radar waves, like light waves, must pass through the atmosphere to travel from one place to another. The characteristics of the medium

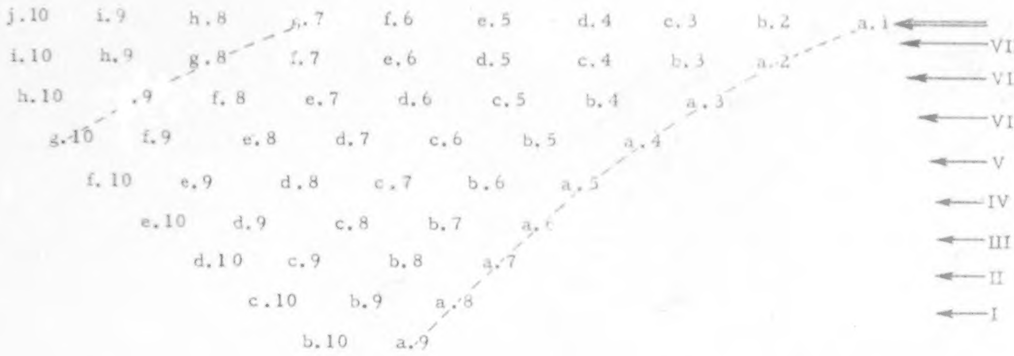

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Figure 16-8.—Generation of precipitation patterns in a wind field.

through which the waves pass affect the manner of their transmission. Thus, although it is sometimes assumed that both light and radar waves follow perfectly straight paths, the properties of the atmosphere are such that the waves are made to follow curved paths.

REFRACTION

The bending of radio or radar waves due to a change in the density of the medium through which they are passing is termed refraction. The measure of the bending that occurs is indicated by the index of refraction from one substance to another. The density of the atmosphere normally changes at a gradual and continuous rate. Therefore, under these conditions the index of refraction would change gradually with increased height. However, under certain conditions, the temperature may first increase with height then begin to decrease (a temperature inversion). More important, the moisture content may decrease more rapidly with height just above the sea. This effect is called a moisture lapse. Either a temperature inversion or a moisture lapse, alone or in combination, may produce a great change in the index of refraction of the lowest few hundred feet of the atmosphere, resulting in greater bending of the radar waves. This may greatly extend or reduce the radar horizon, depending on the direction in which the waves are bent. The radio energy may be so highly concentrated in a limited region of space that it appears to be following a duct. (See fig. 16-9.) The height of the radar antenna with respect to the duct formed by this occurrence is of much importance in producing an effect known as trapping.

In order to locate these trapping layers, the meteorologist uses a refraction index which is called the N curve. The refraction index N is defined as the ratio of the wavelength of an electromagnetic wave in a vacuum to that in the actual atmosphere. The atmospheric pressure, temperature, and moisture content are the parameters used to find the value of N. The bending of radio-radar waves is caused by gradients in the refractive index and not by the value of N itself. In other words, when large changes of pressure, temperature, and moisture content occur with altitude, refractivity must also occur.

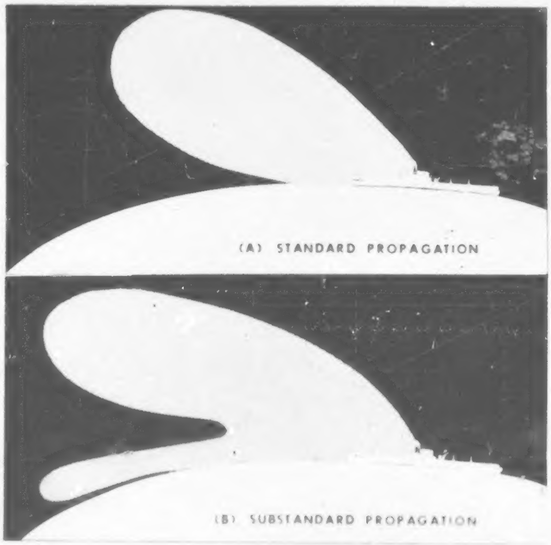

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Figure 16-9.—Normal propagation and surface duct.

Analysis of many meteorological soundings and their associated refraction curves indicates the possibility of classifying these curves into several types. When objects are observed at many times greater than the normal visual range due to the unusual distribution of pressure, temperature, and humidity, we have what is known as superrefractive conditions. When radar waves are reduced, we have subrefractive conditions (quite rare).

The normal N gradient for a standard atmosphere at 60 percent relative humidity is a decrease of 12 units per 1,000 feet. Negative gradients larger than this are defined as superrefractive. However, since radiosonde data is not exact, gradients of N cannot be established precisely, and it is more logical to think in terms of zones. Such zones may be defined as follows:

1. Normal zone—gradients between 0 and −24 per 1,000 feet.
2. Subrefractive—N increases with height.
3. Superrefractive—between −24 and −48 per 1,000 feet.
4. Trapping—gradients larger than −48 per 1,000 feet.

The value of N at sea level ranges from 250 to 450 N units.

CONDITIONS CAUSING DUCTS

What types of atmospheric conditions give rise to the trapping refraction curves and result in extraordinary ranges on VHF and UHF radio equipment? Important meteorological factors that cause changes in the distribution of temperature and moisture throughout the atmosphere are the flow of warm dry air from a land mass out over cooler sea; nocturnal cooling (over land); flow of cool air over warmer sea; and low-level subsidence.

Perhaps the best method by which to consider the question is to subdivide the significant meteorological conditions into seven classifications, discussing their relative effects on radio propagation:

1. Surface (radiation) inversions.
2. Stratum of cool air blowing over warmer sea.
3. Inversions produced by warm continental air blowing over colder sea water.
4. Subsidence inversions in warm high-pressure areas.
5. Coastal stratus type inversions.
6. Frontal inversions.
7. Postfrontal subsidence inversions in cold high-pressure areas.

Surface (Radiation) Inversions

The surface inversions result from nocturnal cooling; that is, loss of heat from the ground by radiation. A temperature inversion is established, and a low-level duct results. This nocturnal cooling is greatest when there is no overcast and, as a result, trapping rarely occurs over land when the sky is cloudy. Radiation ducts will be found chiefly over large flat land areas in the middle and higher latitudes. They will be stronger in the higher latitudes and on the leeward shores of large islands and land masses. One should expect them chiefly at night and in the early morning, when there are clear skies, little or no wind, and low moisture content in the air. When the wind is strong, the skies are clear, and there is little moisture in the atmosphere, the duct will generally disappear early in the morning. Radiation inversions rarely occur more than 1 or 2 miles offshore; and when one exists away from land, the duct will be weakened and the base will gradually rise above the surface of the water.

Stratum of Cool Air Blowing Over Warmer Sea

Cool air blowing over a warmer sea may produce surface ducts. With this particular phenomenon there is no associated temperature inversion, and the entire effect is caused apparently by the evaporation of water into the lower levels of the atmosphere. Ducts usually accompany winds that have blown for long distances over the open seas, their height and size increasing with the wind speed. At higher wind speeds typical of the Pacific trade belt (10 to 20 knots), ducts from 50 to 60 feet high occur quite generally. These ducts are particularly important because they extend for such long distances and because they seem to occur so frequently.

Inversions Produced by Warm Continental Air Blowing Over Cooler Water

When warm air blows over cooler water, a duct is formed by the temperature inversion which is caused by the warmer air aloft. The base of this type inversion is at sea level with the ducting layer extending to several hundred feet. The height of the trapping layer is usually from 200 to 600 feet, and it tends to remain as far as 100 to 200 miles out to sea. Near the coast, the trapping is the strongest just after noon and weakest just after dawn. Fog is generally not associated with this type inversion, but when surface fog appears 5 to 10 miles out to sea, it indicates the presence of a duct.

Subsidence Inversion in a Warm High-Pressured Area

Subsidence inversions are predominant in high-pressure systems. Subsidence results in a temperature inversion which spreads out laterally and creates conditions favorable for a duct.

Such a duct is usually relatively high and may become intense to a height of a few thousand feet. This type of duct will not materially affect surface propagation. The height of such inversion may be estimated from observations of clouds since stratified clouds or thin altocumulus clouds usually form at the base of the inversion during the night or early morning. This type of duct is often found in the Tropics and off the western coasts of continents in the trade wind areas. It is more prevalent in summer than in winter and on the eastern side of the subtropical highs.

Coastal, Stratus Type Inversions

This situation occurs when warm continental air is underrun by cool sea air. Strong surface ducts are the result. These ducts may be expected most often on the western shores of continents in the lower and midlatitudes and where the higher atmosphere is blowing offshore, with the lower cooler portions stationary or possibly even blowing onshore. They vary with the season, being present almost entirely in late spring, summer, and early fall. They are most pronounced during night and early morning.

Frontal Inversions

The temperature increase with altitude in the frontal inversion is a gradual increase and not abrupt, but the moisture content increases sharply throughout the transition zone. The increase of moisture within the inversion layer leads to subnormal rather than trapping conditions. The region in which they are prevalent varies with the season.

Postfrontal Subsidence Inversions

This type of inversion is usually not strong enough to produce more than a slightly less than normal decrease in N with height. The base of the inversion is generally 4,000 to 8,000 feet. This type inversion is most frequently observed in middle and high latitudes and never in the doldrums. They are stronger in winter than in summer, and can increase in intensity and decrease in height at night. Their effect on propagation is similar to that produced by subsidence inversions in warm high-pressure systems.

WAYS OF DETECTING A DUCT

Meteorologists are at present studying carefully the various ways in which weather affects high frequency radio propagation. It is hoped that some day they will have developed a program of forecasting duct conditions by means of which complete and accurate radio coverage may be predicted. Since this goal has not yet been attained, one must use the information at his disposal. Radio propagation for certain frequencies is determined by atmospheric conditions and closely associated with the weather. Consider several rules by which the presence of a duct may be detected. Trapping should be expected when the following conditions occur:

1. A wind is blowing from land.
2. There is a stratum of quiet air.
3. There are clear skies, little wind, and high barometric conditions.
4. In open ocean when a cool breeze is blowing over warm ocean, especially in the tropical areas and in the trade wind belt.
5. Smoke, haze, or dust fails to rise but spreads out horizontally.
6. There are skips in the ground clutter on the scope; for example, when nearby islands are not received on the scope.
7. There is strong fading of the pip. A rapidly fading pip may indicate a duct with ranges on a higher sweep. This rule, however, is not universally true. The duct may be well formed and steady; or when there are standard conditions, changes in the aspect of the target may cause rapid scintillations of the signal.
8. The moisture content of the atmosphere at bridge level is considerably less than just above the sea surface.
9. There is an offshore wind and the temperature at bridge level is 1° to 2°F greater than that of the sea.
10. In open ocean, the temperature at bridge level is definitely less than that of the sea.
11. In trade wind areas, generally for frequencies of 3,000 mc and higher, with low-level antennas.

12. The potential temperature at a level of about 2,000 feet exceeds the surface value by 10°F or more while the mixing ratio for this level is at least 5 gm/kg below the surface value.

NOTE: Keep in mind that potential temperature is determined by moving the parcel dry adiabatically from the 2,000 ft level (in this case) to the surface level.

COMPUTATION OF REFRACTIVE INDICES

Direct measurements of the propagation of electromagnetic energy in a particular medium can be made and, when compared with the speed of light, give a direct measure of the refractive index or refractivity. Because of the complexity of the equipment needed, this approach is impractical for operational use. Therefore, the refractive index must be computed from available radiosonde data. Numerous methods have been devised for computing N. These take the form of tables, slide rules, computers, and nomograms.

The publication entitled "The Analysis and Forecasting of Atmospheric Radar Refractivity," NA 50-1P-1, is issued as an aid to Naval Weather Service Command Units in satisfying an increasing requirement of the operating forces for radar refraction data. It deals with the Arowagram.

NAVAIR 50-1P-7, Computation of Atmospheric Refractivity of the USAF Skew T, Log P diagram deals with the Skew T as indicated by its title.

One version of the Skew T diagram (DOD WPC 9-16-2) has been printed with a grid for determining refractivity already printed on it. Refractivity can be read directly from the traces of temperature and dewpoint entered on the diagram, without the use of an external nomogram. The accuracy and range of computations should cover most of the applications of weather units. Instructions for its use are covered in NAVAIR 50-1P-7.

OCEAN DUCTS AND THEIR DETERMINATION

Evaporation of "oceanic" ducts are not particularly effective in extending the coverage of most shipboard radar because they are usually shallow (75 feet is about average). In order for such a duct to be effective, the radar must be within 50 feet of the sea surface, and its wavelength must be shorter than 30 cm. The normal practice of installing radars on the highest point on the ship places most radars above radar ducts which may be present. For low-sited radars, however, the duct is often useful in extending the normal limit of coverage. In actual practice one can determine from simple meteorological data whether such a radar at a particular wavelength will have extended coverage.

For these purposes a Ductogram was designed to facilitate the prediction of extended radar coverages with an oceanic duct. For input data, the Ductogram utilizes surface observations of air, dewpoint, and sea water temperatures taken on board ship. The quantity of N' is the approximate difference between the refractivity N, at the elevation of the shipboard observation (bridge height), and the refractivity N_s at the water surface. Complete details and illustrations of the Ductogram may be found in NA 50-1P-1.

LOCAL FORECASTS USING MICROANALYSIS

A generalization of the expected weather conditions, which is too inconclusive for many military operations, is the usual result of the large scale analysis that precedes most forecasts. Quite often, and especially during wartime operations, the need arises for a forecast of the exact conditions at a particular locality during a specific time. This kind of forecast requires microanalysis, which should determine whether a particular spot will have cloud coverage at a specific time of the day or night. Local and diurnal effects have to be carefully weighed.

For example, operations might require sending aircraft to their maximum range, making it imperative that they encounter no delays in landing due to weather conditions at arrival time. In this instance the important question is not what type of weather can be expected on a particular day but what type of weather will be expected at a specific time.

To achieve maximum accuracy in this type of forecast, it is necessary to utilize more individual station data than is found on the synoptic chart. Use of small scale weather distribution charts is one method that may be used. Here close scrutinization is made of reported clouds or other weather phenomena. Another useful method is the plotting of information from an individual station on a frequent and continuing basis. Again by following the changes as they occur a more accurate prediction of expected changes can be made.

Both of these methods are extremely useful in tropical areas where the local and diurnal effects are so greatly pronounced.

PRESSURE-HEIGHT FORECASTS

It is important for many practical purposes to be able to determine the altitude at which certain strata of air may exist in the atmosphere, e.g., inversions, isothermal layers, moist layers, etc. The calibration of pressure altimeters also requires the computation of pressure height data.

Pressure-height or pressure-altitude refers to the height of a given pressure surface under standard atmospheric conditions. Computations used in determining "standard atmospheric conditions" within the United States are based on what is commonly referred to as the U.S. Standard Atmosphere. The figures which represent the U.S. Standard Atmosphere are as follows:

1. Mean sea level pressure = 1,013.25 mb
2. Mean sea level temperature = 15°C.
3. Lapse rate = 0.65°C per 100 M to 10,769 M (35,332 ft or 234 mb).
4. Lapse rate above the tropopause = 0.000 with a constant temperature of -55°C (above 35,332 ft or 234 mb).

The actual heights for selected pressure levels based on this standard atmosphere will vary from that in a standard atmosphere as the environmental conditions and elements involved are changed from those of the standard atmosphere.

A pressure altimeter is calibrated to read the heights of points in the U.S. Standard Atmosphere when the pressure scale on the instrument is set at 29.92 inches of mercury. Consequently, if the altimeter is set at 29.92 inches, the indicated altitude of an aircraft will be 4,780 feet at the 850-mb level, 9,880 feet at the 700-mb level, etc. Since the actual atmosphere does not conform to standard conditions, the actual (geometric) heights of the pressure surfaces may depart considerably from that of the pressure heights. For many flight operations this departure of the geometric height of a pressure surface from the pressure height for that surface is of extreme importance.

The distribution of the elements mentioned in the preceding paragraphs may be presented in many combinations on various thermodynamic charts. Each combination of these meteorological variables and their presentation on a particular thermodynamic diagram may be suited to certain specialized types of computation, but no one diagram is convenient for all forecasting uses. However, one diagram which is considered to be especially convenient for problems related to pressure-height computations is called a PASTAGRAM. Detailed information pertaining to this diagram may be found in the booklet entitled "Use of the Pastagram in Pressure-Height Computations," NA 50-1P-501. This booklet also contains tables for determining altitudes based on the U.S. Standard Atmosphere from pressure values. The Pastagram has many meteorological uses, among which are included pressure-height determinations, checking or extrapolating heights in the troposphere, and conversion of "D" value reports to pressure values.

AIR DENSITY AND WATER VAPOR COMPUTATIONS

Air density and the water vapor content of the air have an important effect upon engine performance and takeoff characteristics of aircraft. In this section of the chapter we will discuss methods for computing these elements from a meteorological standpoint. The four most common elements on which the Aerographer's Mate is requested to furnish information are pressure altitude, density altitude, vapor pressure, and specific humidity. All of these may be determined from the Density Altitude Computer contained in NWRF publication

37-0862-066. Complete instructions for the use of this computer may be found in the narrative section of this publication with two examples for computation of density altitude on the reverse side of the computer. Density and pressure altitude are given in feet, vapor pressure commonly in millibars or inches of mercury, and specific humidity in grams per gram or pounds per pound.

DENSITY AND DENSITY ALTITUDE

The density of the atmosphere is a factor in many of the important problems in meteorology, as well as in those related sciences and engineering. The performance of an aircraft or missile depends on the density of the air in which it is flying. It is more difficult to take off from high-altitude airports than from airports at low altitude, and it is more difficult to take off on a hot afternoon than on a cool morning.

The Aerographer's Mate in the field may be called upon to furnish air density information for operational purposes or to give lectures on this subject. In either case, he is confronted with a problem that requires him to spend considerable time reviewing available literature. This is true since synoptic meteorologists do not commonly use air density as a parameter, and thus are not always familiar with the application of density considerations to aircraft operations.

Since both temperature and pressure decrease with altitude, it might appear that the density of the atmosphere would remain constant with increased altitude. This is not true, for pressure drops more rapidly with increased altitude than does temperature. Since standard pressures and temperatures have been associated with each altitude, the density the air would have at those standard temperatures and pressures must be considered standard. Thus, there is associated with each altitude a particular atmospheric density. Remember density altitude is not necessarily true altitude. For example, on a day when the atmospheric pressure is higher than standard, and the temperature is lower than standard, the density which is standard at 10,000 feet might occur at 12,000 feet. In this case, at an actual altitude of 12,000 feet, we have air which has the same density as standard air at 10,000 feet.

Pressure Altitude

The pressure altitude is defined as the altitude of a given atmospheric pressure in the standard atmosphere. The pressure altitude of a given pressure is usually a fictitious altitude. It is equal to the true altitude only in the rare case when atmospheric conditions between sea level and the altimeter in the aircraft correspond to those of the standard atmosphere. Aircraft altimeters are constructed for the pressure-height relationship that exists in the standard atmosphere. Therefore, when the altimeter is set to standard mean sea level pressure (29.92 in.), it indicates pressure altitude and not true altitude.

Density Altitude

The density altitude is defined as the altitude at which a given density is found in the standard atmosphere. If, for example, the pressure at Cheyenne, Wyoming (elevation 6,140 ft) is equal to the standard atmosphere pressure for 6,140 feet but the temperature at that station is 101°F, the density there is the same as that found at 10,000 feet in the standard atmosphere, and an aircraft on takeoff would perform as if the altitude of the runway were at 10,000 feet in the standard atmosphere instead of 6,140 feet.

The effects of air density on the pitot tube and the calibration of the airspeed indicator of an aircraft are such that indicated airspeed and true airspeed are equal only when density altitude is zero; that is, when the air density is that of dry air at standard mean sea level pressure (29.92 in.) and standard sea level temperature (15°C). True airspeed exceeds indicated airspeed when the density altitude increases.

No instrument is available to measure density altitude directly. It must be computed from the pressure (for takeoff, station pressure) and the virtual temperature at the particular altitude under consideration. This may be accomplished by using the Density Altitude Computer or by using a chart in Manual of Barometry, NW 50-1D-510. Remember, virtual temperature is used in the computation of density altitude.

WATER CONTENT OF THE AIR

The capabilities of an aircraft will be affected significantly by the water content of the air.

Water content is frequently described as fog or in the extreme, precipitation. The forecasting of fog and precipitation were discussed quite extensively in chapters 10 and 11 of this manual. Two other frequent requests concerning water content are for vapor pressure and specific humidity.

Vapor Pressure

Vapor pressure is that portion of the atmospheric pressure that is exerted by the water vapor in the atmosphere and is expressed in inches and in tenths of an inch of mercury (Hg). The dewpoint for a given condition depends on the amount of water vapor present, so a direct relationship exists between vapor pressure and the dewpoint. Vapor pressure may significantly affect engine power.

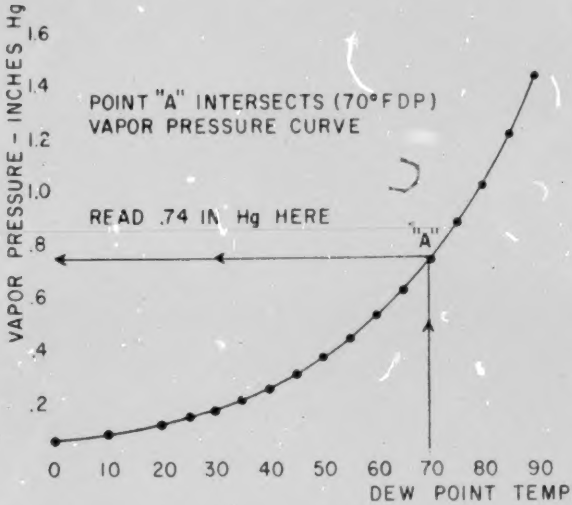
Figure 16-10 shows a chart for determining vapor pressure from the dewpoint temperature. You enter the bottom of the chart with the dewpoint temperature and go up vertically until you intersect the curve and read the vapor pressure at the left side of the chart in tenths of inches and tenths of mercury. This value may also be found on the Density Altitude Computer.

Specific Humidity

The mass of water vapor present in a unit mass of air is known as specific humidity. The mass of the unit of air taken is considered to be a unit mass of moist air. Since the mass of a unit of dry air differs but little from the mass of a unit of moist air, the mixing ratio and the specific humidity are nearly numerically equivalent.

In tropical countries, where temperatures are high and rainfall is excessive, the water vapor content of the air reaches high proportions. All engine test data are calculated on dry air. Since water vapor is incombustible, it is a total loss as far as the engine is concerned. Specific humidity can be determined from the Density Altitude Computer or from the two following methods. The values given in these examples show the amount of water vapor in pounds per pound of dry air:

VAPOR PRESSURE DIAGRAM (30" Hg)



GIVEN: DEW POINT 70° F
FIND: VAPOR PRESSURE INS. Hg

INSTRUCTIONS:
1. DETERMINE DEW POINT TEMPERATURE
2. GO UP FROM DEW POINT TO INTERSECTION OF VAPOR PRESSURE CURVE
3. READ VAPOR PRESSURE AT LEFT OF PAGE

FROM: TABLE 95, SMITHSONIAN METEOROLOGICAL TABLES, NA50-1B-521

AG.724

Figure 16-10.—Vapor pressure chart.

1. If the dewpoint is known, use figure 16-11 (A) and interpolate for the value of specific humidity. In this case the dewpoint temperature is 66°F and the specific humiduty by interpolation is 0.0135.

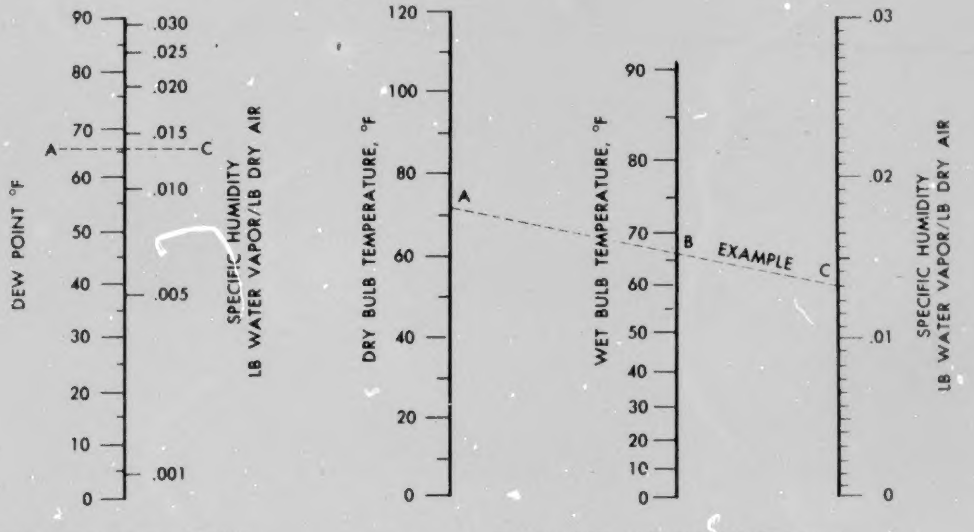
2. If the wet bulb and temperature are known and not the dewpoint use figure 16-11 (B). In this example the dry-bulb temperature is 72°F and the wet-bulb temperature is 67°F. With a straightedge, enter the dry-bulb scale at 72°F (Point A), locate 67°F wet-bulb on the wet-bulb scale (point B), and project a straight line through these two points to the specific humidity scale. It reads approximately 0.0133.

FORECASTING ALTIMETER SETTINGS

Under certain conditions it may be necessary to forecast or develop an altimeter setting for a station or a location for which an altimeter setting is not received. There is also a possibility that an altimeter setting may be required for an area not having a weather station. Recent implementation of the COMET System requires that designated naval air stations prepare and disseminate Plain Language Terminal Forecasts in the PLATFS Code. A forecast of the lowest altimeter setting for the forecast period is required. For these reasons it is important that forecaster personnel have a basic understanding of the importance of correct altimeter settings and a knowledge of a procedure for forecasting altimeter settings. Refer to FMH No. 1 Chapter A8 for computation of altimeter settings.

BASIC CONSIDERATIONS

An altimeter is primarily an aneroid barometer calibrated to indicate altitude in feet instead of units of pressure. An altimeter reads accurately only in a standard atmosphere and when the correct altimeter setting is used. Since standard conditions seldom (if ever) exist, the altimeter reading usually requires correction. It



IF DEW POINT TEMPERATURE IS KNOWN, READ SPECIFIC HUMIDITY FROM CONVERSION SCALE ABOVE

(A)

IF ONLY WET AND DRY TEMPERATURES ARE KNOWN, DETERMINE SPECIFIC HUMIDITY FROM NOMOGRAM ABOVE

(B)

AG.725

Figure 16-11.—Determination of specific humidity from nomogram. (A) When dewpoint is known; (B) When temperature and wet bulb are known.

will indicate 10,000 feet when the pressure is 697 millibars, whether or not the altitude is actually 10,000 feet.

The altimeter is generally corrected to read zero at sea level. A procedure used in aircraft on the ground is to set the altimeter reading to the elevation of the airfield. The altimeter then reads the altitude above sea level and the Kollsman window indicates the current altimeter setting.

Altimeter Errors Due to Change in Surface Pressure

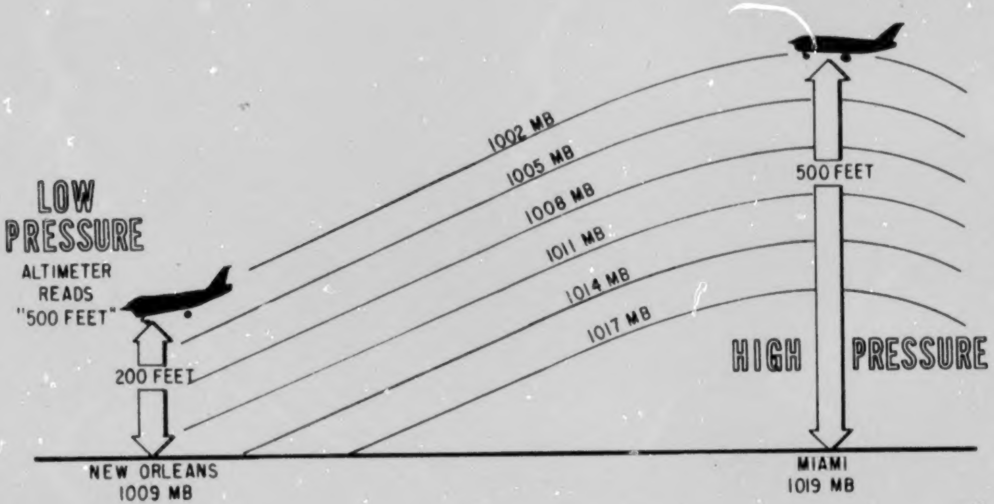
The atmospheric pressure frequently differs at the point of landing from that of takeoff; therefore, an altimeter correctly set at takeoff may be considerably in error at the time of landing. Altimeter settings are obtained in flight by radio from navigational aids with voice facilities. Otherwise, the expected altimeter setting for landing should be obtained by the pilot before takeoff.

To illustrate this point, figure 16-12 shows the pattern of isobars in a cross section of the atmosphere from New Orleans, Louisiana, to Miami, Florida. The pressure at Miami is 1,019 millibars and the pressure at New Orleans is 1,009 millibars, a difference of 10 millibars. Assume that an aircraft takes off from Miami to fly to New Orleans at an altitude of 500 feet. A decrease in the mean sea level pressure of 10 millibars from Miami to New Orleans would cause the aircraft to gradually lose altitude; and although the altimeter indicates 500 feet, the aircraft would be actually flying at approximately 200 feet over New Orleans. The correct altitude can be determined by obtaining the correct altimeter setting from New Orleans, resetting the altimeter to agree with the destination adjustment.

NOTE: The following relationships generally hold true up to approximately 15,000 feet: 34 millibars = 1 inch (Hg) = 1,000 feet elevation. Since 1 millibar is equal to about 30 feet below 10,000 feet altitude, a change of 10 millibars would result in an approximate error of 300 feet.

Altimeter Errors Due to Variation From Standard Temperature

Another type of altimeter error is due to nonstandard temperatures. Even though the altimeter is properly set for surface conditions, it will often be incorrect at higher levels. If the air is warmer than the standard for the flight



AG.726

Figure 16-12.—Altimeter errors due to change in surface pressure.

altitude, the aircraft will be higher than the altimeter indicates; if the air is colder than standard for flight altitude, the aircraft will be lower than the altimeter indicates. (See fig. 16-13.)

FORECASTING PROCEDURE

The first step in the forecasting of altimeter settings is to forecast the sea level pressure for the valid time of the desired altimeter reading. This may be done by using the recommended procedures of prognosis presented in earlier chapters of this training manual. Facsimile prog charts can also be used.

After the value for the expected sea level pressure has been obtained, it must be modified to reflect the diurnal pressure change at the location in question. Pressure tendency charts, locally prepared diurnal curves, and other available information can be used to obtain representative diurnal changes.

The final result of the first two steps will normally be expressed in millibars since it is conventional to work in these units on related charts. If this is the case, then this resultant pressure in millibars must be converted into inches of mercury before it can be used for an altimeter setting.

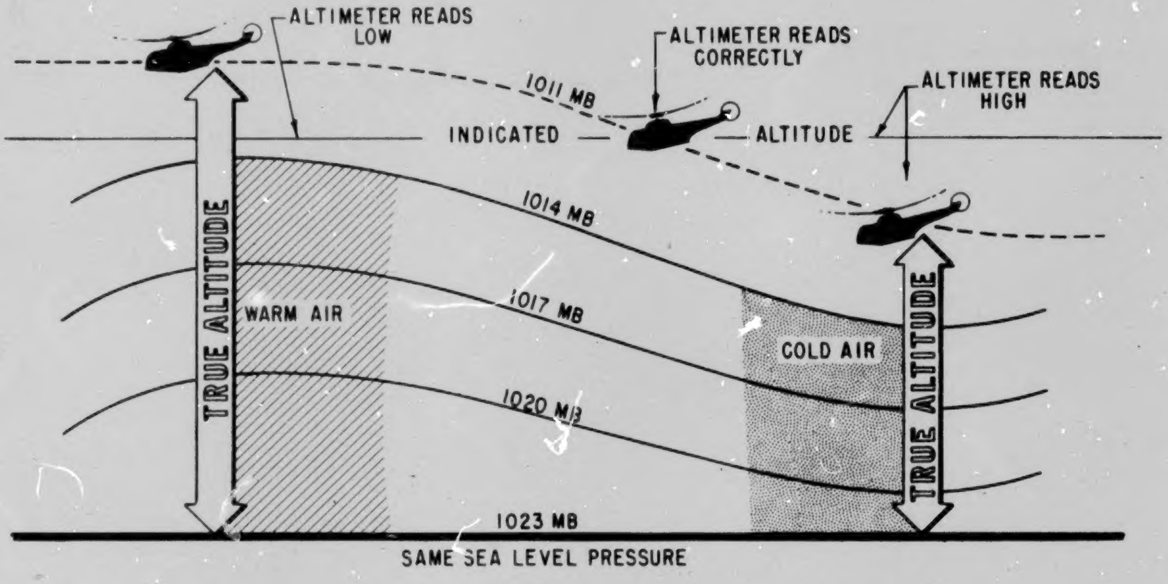
The computations for forecasting in altimeter setting are illustrated as follows:

CASE 1

Forecast sea level pressure	1004.0 mb
Diurnal change	-2.0 mb
Corrected forecast	1002.0 mb
For final forecast, 1002.0 mb converts to	29.95 in.

CASE 2

Forecast sea level pressure	996.0 mb
Diurnal change	+1.0 mb
Corrected forecast	997.0 mb
For final forecast, 997.0 mb converts to	29.44 in.


ALTIMETER READS LOW
ALTIMETER READS CORRECTLY
ALTIMETER READS HIGH
1011 MB
INDICATED
ALTITUDE
1014 MB
1017 MB
1020 MB
1023 MB
TRUE ALTITUDE
WARM AIR
COLD AIR
TRUE ALTITUDE
SAME SEA LEVEL PRESSURE

AG.727

Figure 16-13.—Altimeter errors due to nonstandard air temperatures.

FLIGHT LEVELS VERSUS TRUE ALTITUDE

A flight level is defined as a level of constant atmospheric pressure related to a reference datum of 29.92 inches of mercury. In other words, the altimeter is set to 29.92 regardless of the true altimeter setting, and levels flown thereafter are based on this indication of altitude. For example, flight level 250 is equivalent to an altimeter indication of 25,000 feet when the altimeter is set to 29.92.

In the continental United States and Alaska, FAA regulations require all aircraft to use a standard setting of 29.92 inches at and above 18,000 feet. Also, this standard setting of 29.92 inches is used over the oceans at all altitudes when an aircraft is a distance of 100 miles or more from a defined area.

Over Europe, the standard altimeter setting of 29.92 inches is used at and above 3,000 feet. Pilots must be briefed on low-pressure areas along the flight route. Otherwise, a standard altimeter setting of 29.92 inches could give a false indication that would indicate an altitude significantly higher than the actual altitude. In mountainous terrain, this could prove disastrous unless the pilot is alerted to the presence of these areas of low pressure along the route.

METEOROLOGICAL BALLISTIC FORECASTS

The still (no wind) atmosphere acts on a projectile as a retarding force, whose magnitude is a direct function of the density of the atmosphere. The problem of calculating the extent to which a projectile will be slowed down thus involves measuring the atmospheric density along the projectile's trajectory. This trajectory may be surface-to-surface as illustrated in figure 16-14 or it may be surface-to-air, air-to-air, or air-to-ground.

Although the Aerographer's Mate will most frequently be called upon to furnish corrections for surface-to-surface and surface-to-air trajectories, he might also be asked to provide air-to-ground (bomb) trajectory corrections. The latter are referred to as Q-factors. The magnitude of the retarding effect due to wind and air density is quite important in directing the aim of a gun, since it may reduce the gun's range by more than half of what it would be if the atmosphere were absent.

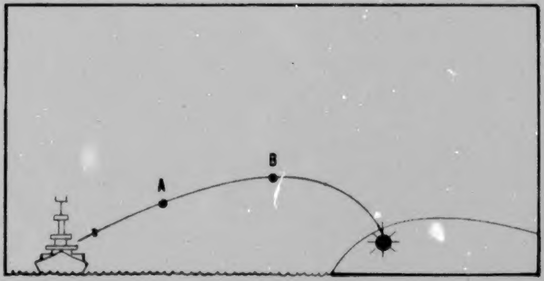


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Figure 16-14.—Example of surface-to-surface trajectory.

Recent developments in new types of naval gun-fired projectiles, such as the rocket assisted projectile (RAP) and improved conventional projectiles have revived interest in meteorological ballistics on the part of the operating forces. The development of simplified procedures for computing and applying ballistic wind and density corrections has become an urgent requirement for fleet operations and for training.

BALLISTIC FORECASTS

The effects of air density and wind upon the trajectory of a projectile cannot be accounted for without considering the meteorological effects. Corrections for wind and density must be applied in the form of "ballistic" wind and density corrections, for which additional computations, requiring some knowledge of meteorology, are necessary. These computations demand measurements, equipment and skills usually available only to Naval Weather Service Command units, and instructions for making them are written primarily for use by meteorological personnel. The dissemination of ballistic wind and density forecasts for Fleet operating units is the responsibility of the Naval Weather Service Command.

The areas for which ballistic forecasts are issued include most of the northern hemisphere and a significant portion of the southern hemisphere. Requests for ballistic wind and density forecasts should specify the location and time period of interest and should be forwarded to the cognizant Fleet Weather Central in accordance with NavWeaServComInst 3140.1 (Series).

NATO STANDARD BALLISTIC METEOROLOGICAL MESSAGE

The United States, by ratification of NATO Standing Agreements (STANAGS), has adopted the standard atmosphere and the standard ballistic meteorological message formats approved for use by forces of the North Atlantic Treaty Organization (NATO). All ballistic wind and density forecasts are transmitted in the NATO format unless specifically requested otherwise.

The NATO Standard Ballistic Meteorological Message is applicable to 3-inch through 8-inch standard projectiles. It consists of two parts: (1) a mandatory preamble, followed by (2) six figure groups of ballistic data, in pairs, which, for naval gunfire, may number up to ten or more pairs. The format and content are contained in the technical manual Ballistic Wind and Density for Naval Gunfire, NavOrd OP 3784.

DETERMINATION OF BALLISTIC ELEMENTS

With the introduction and advancement of computer technology many of the ballistic corrections required for missiles and naval gunfire are applied automatically. However, raw data must still be determined and it is necessary for the AG to familiarize himself with the computational procedures involved.

Ballistic Density

Ballistic density is obtained by computing the weighted average of the relative values in the various zones, the relative values in each zone being obtained by first computing the zone density from the mean virtual zone temperature and the pressure at the mid-point of the zone, and then expressing this zone density as a percentage of the standard density for the zone. For example, if the ballistic density for a particular zone is expressed as 105 percent, then the actual atmospheric density is 5 percent greater than the standard ballistic density for that height.

U.S. Navy range tables (with the exception of RAP) and fire control computer cams are based on the Navy Ballistic Standard Atmosphere which is described as a density of 1.2034 kg/m^3 at a pressure of 1,000 mb with a temperature of 15°C and a relative humidity of 78 percent. This data differs from the NATO atmosphere in that the Navy atmosphere is lower up to 7,000 meters. At approximately 8,500 meters the values of density in the two atmospheres are equal, but the Navy atmosphere rapidly becomes the more above that level. The physical procedure for computing ballistic density is contained in NavOrd OP 3784, mentioned earlier in this section.

Ballistic Wind

Ballistic winds for conventional gunfire are obtained in much the same manner as ballistic densities. The same weighting factors are used for both range and crosswind components. For rocket-assisted projectiles, however, wind weighting factors are not the same for range wind as for crosswind. Thus, a correct determination of the range wind component depends upon a ballistic wind based on range wind weighting factors, and determination of the crosswind component depends upon a ballistic wind based on crosswind weighting factors. Both components must be used to obtain the ballistic wind. Ballistic wind computational procedures are contained in NavOrd OP 3784.

Q-FACTORS

The effects of wind are as important to the fall of a bomb as they are to the trajectory of a projectile. In the case of bombing, however, the effects of wind are treated slightly differently than the effect of wind on a projectile, primarily because the wind corrections must be of such a nature that they are readily adaptable for use as an input to the bombsight. The term "Q-factor" is subsequently used and is a modification of the ballistic wind concept. The procedure for computing Q-factors may also be found in NavOrd OP 3784.

SOLAR RADIATION-HIGH ALTITUDE (SOLRAD-HI) OBSERVATIONS

The SOLRAD-HI program which is currently under development involves the launching of a series of satellites for the measurement of high altitude solar radiation emitted by the sun.

TECHNICAL FACTORS

Military activities concerned with satellite surveillance, communications, and research and development are becoming increasingly concerned with solar-geophysical activity. Since the Aerographer's Mate is directly involved in satellite surveillance for the receipt of meteorological data and involved either directly or indirectly in communications, he should keep current on certain factors related to these fields. In the case of satellite surveillance, solar geophysical activity has been determined to affect the amount of drag on some space satellites; cause some changes related to orbital parameters; and produce various tracking difficulties. In the case of communications, the effects are primarily in the area of the quality of transmission and reception of data. There are many technical factors involved in the field of solar geophysical study which are beyond the scope of this Rate Training Manual. However, the AG should have a basic understanding of some of the characteristics related to this field because, due to his involvement in receiving satellite data, he may be called upon at a later date to perform functions related to the acquisition of solar geophysical data.

Some of the physical features of the solar disk were presented in chapter 3 of this manual.

EFFECTS OF SOLAR FLARES

Outbursts of radio noise often accompany solar flares, particularly the larger ones. This noise may take the form of a sudden onset of radio noise in a narrow band at high frequency, drifting downward in frequency to below 100 MHz over a period of from 2 to 5 minutes or in some cases over a period of a few seconds. Flare associated radio disturbance may occur as an onset of noise over a band of several hundred MHz commencing during visible phases of a flare and lasting anywhere from a few minutes to several hours after the flare ceases to exist.

Solar flares are found to have a direct relationship with disturbances occurring in the earth's ionosphere. These ionospheric disturbances occur coincident with and some hours following flares and solar geomagnetic storms, indicating that flares emit ultraviolet radiation, x-rays, and electrically charged particles. The more intense magnetic storms are strongly correlated with solar flares. However, it must be kept in mind that although ionospheric disturbances may be correlated with solar flare activity, all solar flares do not necessarily create ionospheric disturbances.

During the disturbances of the ionosphere attributed to solar flare activity, long distance radio communications may deteriorate or be completely blacked out for several hours. Disturbances may create changes in the strength of the geomagnetic field resulting in interruptions in land line currents sufficient to affect wire communications.

SOLAR PROMINENCES.—Solar prominences appear in various configurations. Among these are those designated as "loop" prominences. The loop prominence is indicative of a highly eruptive solar activity center in its most active phase. When loop prominences are observed on the eastern side of activity centers, solar flares such as those mentioned in the preceding paragraph may be expected during the following 2 weeks. Solar prominences are discussed in more detail in chapter 3 of this manual.

SOLAR OBSERVATIONS

Solar observations for the purpose of studying solar emissions are currently performed in a variety of ways, among which are included the following:

1. Optical observations. Optical observatories, through the use of visual telescopes, provide detailed observations for analysis, prediction, and detection of solar activity. They also provide world-wide, 24-hour coverage of the sun's physical appearance. Although much valuable data is obtained in this manner, the limitations placed on this means of observation are readily apparent.

2. Radio observation. These observatories supplement the optical observatories through the utilization of radio telescopes. These telescopes in actuality are large antenna-receiver combinations. They monitor solar radio emissions at discrete (isolated) frequencies set aside for this purpose. They provide solar maps in selected wavelengths for detailed analysis.

3. Satellite observations. This method of solar observation currently provides observations of solar x-ray radiation, electromagnetic particles, and solar wind characteristics. Satellites also provide early warning of changes in the interplanetary or near-earth space environments. They also provide increased solar coverage since they may be positioned to detect solar events which occur on some portions of the sun not visible from the earth.

ARCTIC AND ANTARCTIC ICE OBSERVATIONS

The study of the air-ocean environment, especially in polar meteorology, would be incomplete without the presentation of some information related to the characteristics of sea ice.

The U.S. Naval Oceanographic Office (NAVOCEANO) was engaged in supporting fleet operations within the arctic and antarctic seas through the implementation of an ice observation and forecasting program. This responsibility has now been assumed by the Naval Weather Service Command and assigned to Fleet Weather Facility, Suitland. Included in this program is the issuance of bulletins which consist of compilation of data obtained from shipboard, shore, and aircraft observers who forward their ice observations to FLEWEAFAC, Suitland. These observations may be received as special reports or they may be appended to routine weather reports. For this program to be fully effective, it is essential that all vessels and aircraft operating in ice areas cooperate with the NAVWEASERVCOM. Each bit of information adds to the steadily increasing knowledge of these least traveled and remote seas. Such data also add to the accuracy and usefulness of the global sea ice forecasts produced and distributed by FLEWEAFAC, Suitland.

Greater utilization of Arctic and Antarctic areas has brought about a need for Aerographer's Mates to develop an understanding of sea ice, the terminology used, its general distribution, and the Ice Observation Program. The need to protect our northern borders from attack has made it necessary to establish a network of listening posts such as radar sites, throughout the Arctic. These weather stations and radar sites must be supplied. Usually these supplies are brought by ships. These ships must be aware of where the ice lies in relation to their position. Only in summer does the icepack retreat from our northern coastlines. Therefore, this is the season when supply routes must be traversed.

For these and other reasons personnel are given thorough training in all aspects of ice observing prior to assignment to operational duties. These duties may include assignment to aerial ice observing units or to various arctic or antarctic bases.

ICE IN THE SEA

Ice in the sea consists, for the most part, of either sea ice formed by the freezing of top layers of the ocean, or icebergs originating from glaciers or continental ice sheets. Sea ice accounts for around 95 percent of the area ice encountered, but bergs are important because of the manner in which they drift from their point of origin, constituting a navigation hazard. A certain amount of ice encountered at sea originates in rivers or estuaries as fresh water ice; however, as it is already in a state of deterioration by the time it reaches the open sea, its importance is local.

ICE OF LAND ORIGIN

Ice of land origin in the sea, though often spectacular, is of minor importance in Arctic operations, except in localized areas. Icebergs are large masses of ice detached from the fronts of glaciers, from glacier ice tongues, or from the shelf ice of the Antarctic. Smaller masses, termed growlers and bergy bits, may originate from glaciers, or may be formed from the disintegration of icebergs and other masses of land formed ice.

FORMATION OF SEA ICE

When sea water with a salinity that is higher than 25.0 o/oo is cooled at the surface, the density of the surface layer increases, causing convective movements that continue until the surface water is cooled to the freezing point and ice begins to form. Elongated crystals of pure ice are produced first. Since the salinity of the surface water increases with cooling, the convective currents are maintained. With continued freezing, the ice crystals form a matrix in which certain amounts of sea water become trapped. The more rapid the freezing, the greater the amount of sea water that will be enclosed in the ice. When the temperature of the ice is further lowered some of the trapped sea water freezes, making the cells containing the salt brine smaller, resulting in the salt concentration in the enclosed brine becoming greater. Thus, sea ice is made up of crystals of pure ice separating small cells of brine whose concentration of salt depends on the temperature of the ice. If the ice is cooled to low enough temperatures, solid salts will crystallize out.

With a rise of the temperature of sea ice, the ice which encloses the brine-filled cells melts and the salt crystals begin to dissolve. As melting progresses, the brine cells become larger; as the temperature approaches 0°C, the cells join, permitting the trapped sea water to flow downward. Where the sea ice has become hummocked (ridge of ice), most of the salt brine will flow downward leaving only pure ice. This ice can be used as a source of potable water. Sea ice which has not been hummocked will become soggy and disintegrate.

Salinity of ice depends on how fast it freezes, its age, and on the temperature changes to which it has been subjected. When sea ice freezes very rapidly, brine and salt often accumulate on ice surfaces, making the surface wet at temperatures as low as -30° to -40°C. This accumulation greatly increases friction on sled runners and skis.

Sea ice properties differ greatly from those of fresh water ice. The properties of sea ice are dependent upon the amount of enclosed brine and the number of air bubbles left in the ice if all or part of the brine has trickled down.

The density of sea ice is dependent upon its content of brine and air bubbles. The density of pure ice at 0°C is 0.9168, but the density of sea ice may be either above or below that of pure ice, depending on its brine and air bubble content.

GENERAL TERMINOLOGY

To make an intelligent approach toward an understanding of ice, it is first necessary to know some of the more or less standard terms by which ice is observed and reported. This terminology is a new language. It is very broad and is used almost exclusively by those who work with ice. For further information on terms used for reporting ice, consult Ice Observation H.O.Pub.No. 606-d. Some of the most commonly used terms are listed in the following sections.

Sea Ice

Sea ice is that ice formed by the freezing of sea water at temperatures in the neighborhood of -2°C, subject to certain conditions, as stated in the previous section on the formation of sea ice. The first sign that the sea surface is freezing is an oily opaque appearance of water. This appearance is caused by the formation of spicules, minute ice needles, and thin plates of ice known as frazil crystals, which increase in number until the sea surface attains a thick, soupy consistency. This is known as grease ice. Upon further freezing, and depending on wind exposure, waves, and salinity, the grease ice develops into nilas, an elastic crust with a matte surface, or into ice rind, a brittle shiny crust. Except in wind-sheltered areas, the slush, as it thickens, breaks up into separate masses, and is frequently characterized by a pancake form. The raised edges and rounded shapes result from collisions of the cakes. With the continuation of low temperatures, the cakes freeze into a continuous sheet.

Age

The age of the ice is defined as the stage in the ice cycle from inception to dissolution. There are basically three stages in the formation of sea ice. YOUNG ICE is newly formed level ice. Its thickness is from 10cm to 30cm (4 in. to 12 in.). Young ice is further classified as gray ice and gray-white ice. FIRST-YEAR ICE is more or less unbroken level of ice of not more than one winter's growth, originating from young ice. Its thickness is from 30 cm to 2 m (12 in. to 6.6 ft). First-year ice may be subdivided into thin first-year ice, medium first-year ice, and thick first-year ice; the latter is more than 120 cm (4 ft) thick. OLD ICE is extremely heavy sea ice which has survived at least one summer's melt. Old ice may be subdivided into second-year ice and multiyear ice.

Topography

Topography, or the configuration of the sea surface, deals with the degree of surface roughness of sea ice from flat to extremely rough. The terms most frequently used to describe the topography of the ice, sometimes used in combinations are listed below.

RAFTED ICE.–Rafting is associated with young ice and young first-year ice. Rafting occurs when ice cracks or floes are forced together due to the pressure of wind and is formed by one cake overriding another. Rafted ice has well-defined contours and when observed, may be regarded as a relatively recent occurrence.

RIDGED ICE.–Ridging is associated with first year ice. It is another form of deformed ice in the form of a ridge or many ridges. It is usually much rougher than rafted ice.

HUMMOCKED ICE.–Hummocking is associated with old ice. It is defined as ice piled haphazardly into mounds or hillocks. At the time of formation, hummocked ice is similar to rafted ice, except that the former requires a greater degree of pressure and heaping than the latter.

Water Features

There are a great variety of water features associated with sea ice. Some of the most common features are as follows. A FRACTURE is any break through the ice. LEAD is a long, narrow, but navigable fracture or passageway in sea ice. The lead could be open or refrozen. PUDDLE is a depression on sea ice that is usually filled with melted water caused by insolation when they have burnt completely through the ice. They are then called THAW holes. POLYNYA is any sizable sea water area that is enclosed by sea ice–simply expressed, a large hole in the ice.

Size

Generally, sea ice is categorized into seven different sizes. The sizes are given in the following section. Refer to figure 16-15 for relative sizes and a comparison to other features.

ICEBERGS

Icebergs are large masses of floating (or stranded) ice derived from the fronts of glaciers, from glacier tongues, or from the shelf ice of the Arctic/Antarctic. They are products of land, and not of the sea. Their structure, and to some extent their appearance, depends upon the source from which they are derived. Arctic bergs originate mainly in the glaciers of Greenland, which has 90 percent of the land ice of the north polar region. Arctic bergs are irregular in form and take many varied shapes. Most common are the irregular cone-shaped bergs, produced by glaciers that have plowed across the uneven foreland on their way to tidewater. They differ entirely from the flat-topped straight-sided bergs (tabular bergs) originating where the ice sheet itself is thrust directly out to sea. (See fig. 16-16.)

Irregular icebergs known as glacier bergs are often accompanied by rams which are a protrusion from the submerged portion of the iceberg. These rams can be a great hazard to vessels that might pass close to an iceberg.

The ratio of mass of the submerged portion of a berg to its total mass is equal to the ratio of the specific gravity of the berg to the water in which it is floating. On account of the origin of glacial ice in compacted snow, berg ice contains up to 10 percent trapped air and is, therefore, somewhat less dense than ordinary ice. It is often erroneously assumed that a berg that is one-eighth above water and seven-eighths submerged should be floating with a draft seven times its height above water. These ratios hold

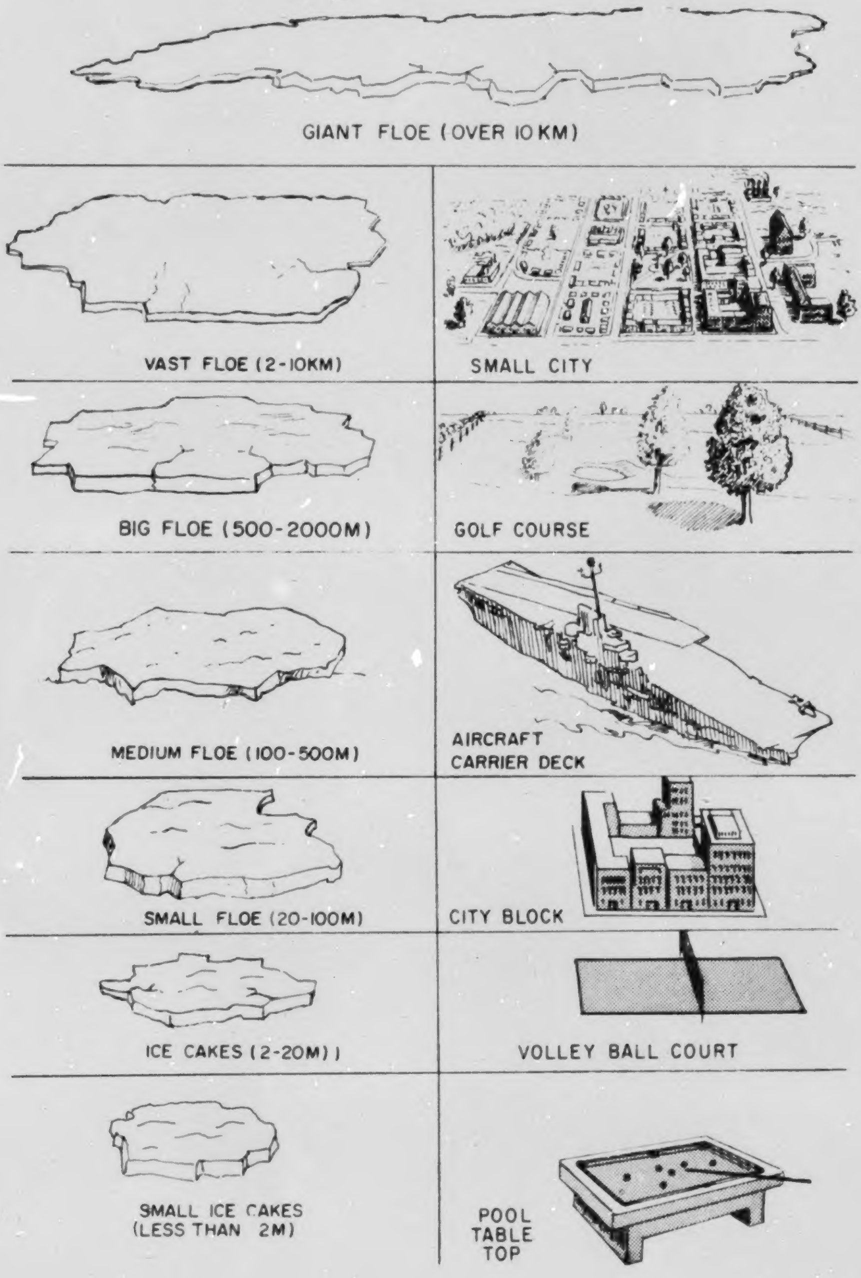
GIANT FLOE (OVER 10 KM)
VAST FLOE (2-10KM)
SMALL CITY
BIG FLOE (500-2000M)
GOLF COURSE
MEDIUM FLOE (100-500M)
AIRCRAFT CARRIER DECK
SMALL FLOE (20-100M)
CITY BLOCK
ICE CAKES (2-20M))
VOLLEY BALL COURT
SMALL ICE CAKES (LESS THAN 2M)
POOL TABLE TOP

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Figure 16-15.—Sizes of sea ice.

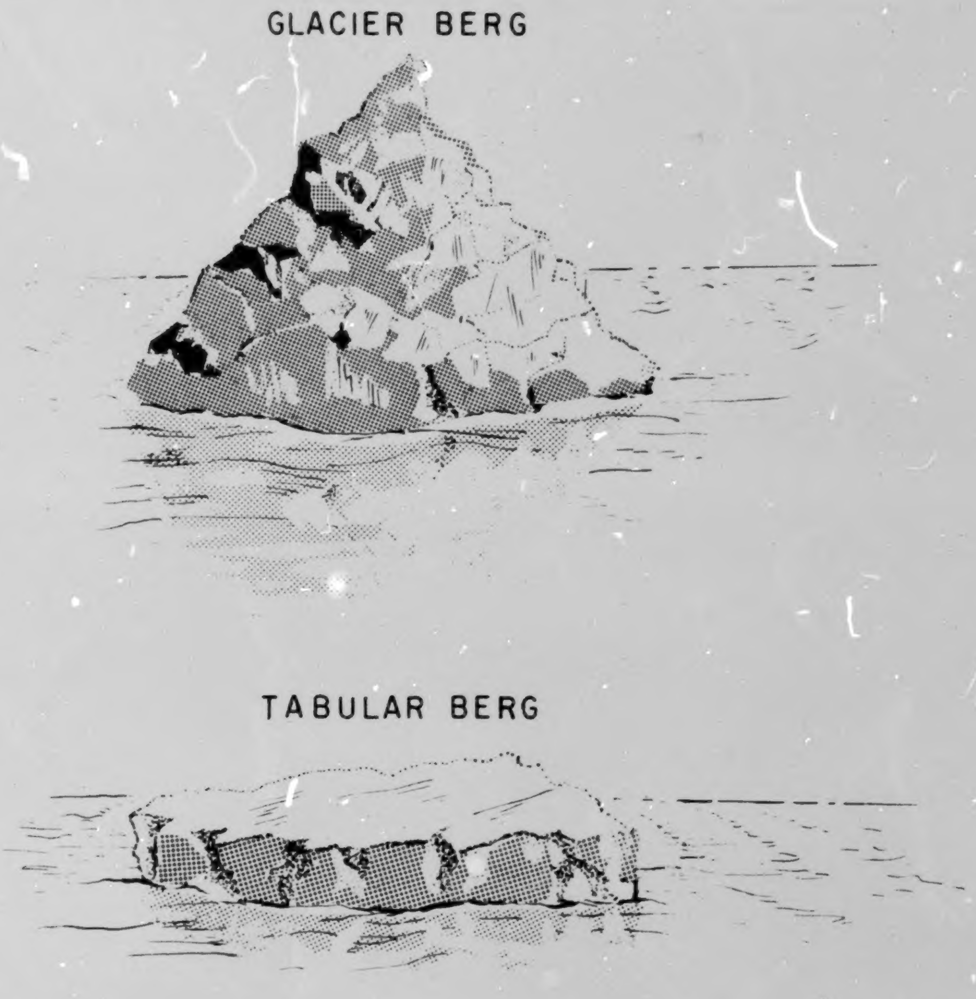

Figure 16-16.—Types of icebergs.

ood only for mass, and not for linear dimenions. Actual measurements of Arctic bergs show hat the draft is seldom more than five times the xposed height for the blockiest bergs and may e as low as one or two times the height for the innacled and irregular types.

BERGY BITS and GROWLERS, like icebergs, originate from glaciers or are formed from the disintegration of icebergs and other masses of land ice. A BERGY BIT is a medium sized fragment of glacier ice and is about the size of a small cottage. A GROWLER is a small fragment of ice awash, smaller than a bergy bit, usually of glacial origin, and generally greenish in color. It is about the size of a grand piano.

MOVEMENT OF ICE

Sea ice, other than fast ice (all types of ice, either broken or unbroken, attached to the shore, beached, stranded in shoal water, or attached to the bottom of shoal areas), in sheltered bays or along the coast is continually in motion as a result of the effects of wind, tide, and current. Although this motion may be the same for a time over a considerable area, there are a number of factors tending to produce differential motion of adjacent masses. Cakes, for example, vary in area and thickness, so that the effect of wind and current differs on different masses of ice. Wind and current are also subject to continual local variations—wind from the usual meteorological causes, and current from tidal effects.

The swinging or turning of floes is due to the tendency of each cake to trim itself to the wind when the pack is sufficiently open to permit freedom of movement. In close pack (any large area of floating ice driven closely together), this tendency may be produced by pressure from another floe; but since floes continually hinder each other, and the wind may not be constant in direction, even greater forces result. Thus, wind produces rotation as well as translation. This screwing or shearing effect results in excessive pressure at the jutting corners of floes, and forms a hummock of loose ice blocks. Ice undergoing such movements is called pressure ice and is extremely dangerous to vessels.

In its motion, the ice opens and shuts like an accordion; a number of leads must be present, otherwise the ice could not move. In summer these leads remain open, except in very high latitudes, but in winter they are soon frozen over with young ice. Swell also tends to break up ice, as well as the vertical movement of the tide in narrow or shallow waters. As a result of these factors, the ice is alternately being broken up, even throughout the winter, and subjected to pressure. The onset of pressure or release of pressure may happen at any time of year, even during the lowest midwinter temperatures.

DRIFT OF ICE

While the general direction of the drift of icebergs over a long period of time is known, it may not be possible to predict the drift of an individual berg at a given place and time, for bergs lying close together have been observed to move in different directions. They move under the influence of the prevailing current at the depth to which they are submerged, which often may be in opposition to the existing wind and sea or surface drift.

Pack ice drifts with the wind and tide usually to the left of the true wind in the Southern Hemisphere and to the right in the Northern Hemisphere. The speed of drift may not depend entirely upon the strength of the wind, since it is influenced greatly by the presence or absence of open water in the direction of the drift, even though the open water is somewhat distant.

Neglecting the resistance of the ice, Ekman's theory of wind drift calls for the ice to drift 45° from the wind direction. Observations show that the actual drift is about 30° from the wind direction on the average, or very nearly parallel to the isobars on a weather map. In winter, when the ice is more closely packed and offers more resistance, its drift deviates less from the wind direction than in summer, and tidal influences become more important.

The speed of drift of pack ice can be fairly closely determined from the wind speed. Observed average speeds of drift of ice in the Northern Hemisphere ranges from 1.4 percent of the wind speed in April to 2.4 percent of the wind speed in September.

The general circulation of the ice of the Arctic Ocean is determined by the direction of the ocean currents, which are the result of two chief factors: the circulation of the atmosphere above the polar basin and the surrounding adjacent seas, and the influx into the polar basin of water of oceanic and river origin, with a compensatory outflow of the water from the polar basin.

BREAKUP AND MELTING OF ICE

Heating agents

Ice and snow are evaporated and melted by direct absorption of radiation and by conduction of heat from the surrounding air, rocks, or water. The ultimate source of heat energy is the sun in either case, but the relative importance of radiation and conduction in melting ice will vary with climatic conditions in different localities.

In the case of the longer wave lengths in the infrared portion of the spectrum, which makes up slightly over half the total radiant energy received from the sun, the proportions reflected by a snow surface are only 15 to 25 percent. The proportions reflected by ice and water are correspondingly less; and since water is opaque to infrared radiation, the heat absorption by water in this region of the spectrum is concentrated in the surface layers where it is of most significance to the melting of ice. It is obvious, therefore, that a surface interrupted with areas of water, either leads between floes or pools of melted water accumulating on top, will absorb much more radiant heat than a continuous ice or snow surface. Once disintegration of an ice sheet has proceeded to the point where free water surfaces appear, the rate of further disintegration is accelerated.

Evaporation

The absorption of heat by ice or snow results in either evaporation or melting. Melting takes place as soon as the temperature of any superficial layer of the ice surface is raised above the freezing point, but evaporation may occur at any temperature. Under still air conditions, the layer of air nearest the ice will soon become saturated with water vapor. Evaporation therefore depends on the diffusion of water vapor from the surface and generally proceeds at a relatively slow rate. Air currents tend to increase the rate of evaporation by inducing turbulent mixing of the layers of air near the ice and by bringing new unsaturated air masses from drier regions. Under conditions of low relative humidity, brisk wind may therefore result in ablation (removal for cause) of large quantities of ice and snow even though the air temperature never reaches the melting point and no melting occurs.

Melting

Melting of ice takes place mostly at the expense of the heat of the surrounding water. This heat may have been absorbed from solar radiation in the vicinity or provided by currents originating in warmer latitudes. Melting also results from direct absorption of radiation by the ice and from contact with warm air. Ice will condense dew from warm, moist air on its surface, and each increment of moisture so condensed will melt several times its weight of ice in the ratio of the latent heat of evaporation to the heat of fusion.

Another factor tending to accelerate the rate of ice melting from solar radiation, once it has started, is the increased stability of the surface layers of the sea brought about by the freshening effect of the melt water. Mixing between the surface and deeper layers, already diminished by the wave-damping action of floating ice, is further decreased by the formation of a surface stratum of relatively low density. In regions where the spring melting of ice is brought about chiefly by atmospheric transfer of heat from lower altitudes, and where local fogs restrict the solar radiation reaching the ice and sea surface, the fresh surface layers of sea water may become greatly chilled, and the rate of melting is reduced. Here, vertical exchange of water caused by wind, sea, current and tides contribute heat to the upper layers and expedite clearing of ice.

Stages of Disintegration

In spring, as the duration of daylight begins to increase and the mean air temperature at the sea surface rises, the snow cover of the sea ice and the top layers of the ice begin to thaw. Under conditions of low humidity, most loss on the upper surface of the ice takes place through evaporation imperceptible to the ordinary observer. Where humidity is higher, pools of dew and melt water form on the surface. This fresh water, running down through cracks and holes in the ice, freezes again on contact with the cold sea water, thus sealing the openings. However, when cracks extend only part of the way through the ice, they are widened by the expansion of this water freezing in them; and even though plugged at the top, they will now extend through to the water. On further rising of the air temperatures and melting of the surface, these cracks open up again, and fresh water in a layer as much as to 2 to 3 feet thick flows under the ice.

Decay of the pack is accelerated by mechanical attrition from the swell. The physical erosion of the floes produces scaling, resulting in the formation of a quantity of small blocks and brash. The scaling process enables the sea to reach more extensive areas of ice where the reduction to finer and smaller particles takes place.

The final stages of melting vary with the type of ice. Ice of one winter's growth melts readily in low latitudes, if brine is still present. The internal melting due to variation in the salt content produces a honeycombed appearance with a much greater surface area. Since the rate of heat absorption through conduction is proportional to the area exposed, the rotten ice so formed quickly disappears. Fresher and hummocky ice is longer lived. The old floes are heavily undercut at the waterline, but honeycombing is rare, owing to the absence of salt. The years-old hummocks of the Arctic pack, having a homogeneous structure of nearly salt-free ice, and having a minimum of exposed surface in proportion to their bulk, survive the longest in warmer waters.

Breakup on rivers usually occurs 3 or 4 weeks after the mean air temperature has risen above 32° F. Ice on lakes breaks up 2 or 3 weeks later, and sea ice may break up at about the same time. The breakup on the rivers is both a rapid and violent event. The force of the water flooding down the channels tears up the ice and drives it seaward at a rate of around 4 knots. Bends or constrictions in the channel cause temporary piling up of the ice with the result that the water and the blocks flood the valleys until pressure breaks the ice barriers. In a week or less, the entire river rids itself of ice. River levels rise tremendously during the breakup, reaching heights of 70 feet or more over the winter levels.

EFFECT OF ICE ON NAVAL OPERATIONS

This discussion is tailored to the requirements for Arctic operations, since we are primarily concerned with the effect ice has on resupply of Arctic bases and the feasibility of subsurface operations in the Arctic.

As has been pointed out previously in this chapter, bases have been established and they must be resupplied. This occurs mostly during the summer months of June, July, and August. Operations into the Arctic regions must be thoroughly planned and prepared in advance to insure their success. A knowledge of some of the requirements will aid you in assisting in these plans and preparations. You are primarily concerned with furnishing information as to the structure dissipation, movement, and breakup of the ice.

Before operations are undertaken in the Arctic, elaborate planning such as in any naval operation, is done in great detail. In the Navigation and Meteorology Annex to a standard Arctic operation plan, pertinent information should be outlined on the navigable sea ice areas as well as instructions for determining and forecasting movements on the area of operation.

From your knowledge gained through the study of sea ice in this chapter, and through further study of available publications, you can readily see how the meteorological office can be of invaluable assistance in planning a successful operation into the Arctic regions. To accomplish this task you must be aware of and alert for sea conditions which could favorably or adversely affect the operation.

OPTIMUM TRACK SHIP ROUTING (OTSR)

The international steamer tracks in use today are modifications of routes laid out by LT Matthew Fontaine Maury, USN (1850). Among other things, these routes were designed to optimize steaming distance and reduce the probability of encountering delaying or damaging storm conditions.

In earlier years, the climatological or statistical approach was the best method available to mariners for prevoyage route determination. Although this method avoids areas of known high frequency of storms, blind adherence frequently results in low efficiency. Conditions far from average may exist along a seasonal route, or portion thereof, such that an out-of-season track is often a much better route.

Today a thoroughly tested ship routing service, based upon considerations of extended weather forecasts, is now available. To date, the Navy Oceanographic Office, which developed the program, the Fleet Weather Central at Norfolk, Virginia, and the Fleet Weather Central at Alameda, California, have successfully undertaken hundreds of routings.

BASIC PRINCIPLES AND OBJECTIVES

Optimum Track Ship Routing (OTSR) is an advisory service which provides recommended ship tracks to minimize time enroute and to reduce storm damage to ships and cargo while at the same time considering any special requirements of the ship. The procedure can be tailored to assist in accomplishing any mission with a higher degree of efficiency on an average.

In essence the program utilizes an effective system of long range forecasting to predict the fastest and safest port-to-port route for transoceanic voyages. This system does not promise smooth seas, sunny skies, and light winds, but rather the high probability of sailing the most rapid obtainable route with safety features unobtainable on nonrouted voyages.

The OTSR service is normally of significant benefit only for voyages where the distance from port-to-port exceeds 1500 miles and there is a choice of routes. A route in which geography dictates the track would not lend itself to OTSR.

The program can be broken down into three phases: route selection, surveillance, and summary.

Route Selection

When a Naval Weather Service unit having responsibility for providing OTSR service receives a request for a route, sophisticated techniques are applied to all available data. Weather prognoses, sea condition prognoses, ocean current data, ship and cargo characteristics, etc., are all utilized to determine the recommended route.

When the route has been selected and tested, it is sent by dispatch directly to the ship, arriving from 24 to 48 hours before departure.

Surveillance

From this point on, a continuous watch is kept on changing conditions which might affect the recommended route. The individual attention rendered to each ship while underway is one of the most important aspects of the program. Each day the routed ship sends a brief message which includes the position, course, speed, wind, and sea conditions. When weather deviations from the original forecast are detected, the ship's route can be modified to avoid undesirable weather conditions. Diversions to the original route are not the rule but do occur, particularly in winter, especially in regions where changes in pressure patterns are apt to be sudden and of considerable magnitude. The number of recommended diversions is surprisingly small. To date, the average is one per voyage with a maximum of three.

Upon completion of a voyage, a voyage summary memorandum is sent to the weather service which provided the routing service. This report indicates position of ship, sea, swell, and wind conditions; and ship's heading and speed at specific times. This summary provides a means for determining the effect of wind and sea conditions on the sailing and for continuing development of ship routing techniques and evaluation of the program.

NAVAL WEATHER SERVICE RESPONSIBILITIES

Optimum Track Ship Routing services are available for all U.S. Navy ships in accordance with established criteria. The Fleet Numerical Weather Facility at Monterey provides this service for the entire Pacific Ocean while the Fleet Weather Central provides service for the Atlantic Ocean. Additional information is available in NAVWEASERVCOMINST 3140.1 (Series).

BENEFICIAL RESULTS

The chief benefits of this service are to allow for a minimum time en route which affords a savings of fuel oil, increased passenger comfort, the reduction of damage to the ship or cargo, and the avoidance of rough seas and severe storm areas consistent with time requirements. You can readily see that a program of this type would also be beneficial to commercial vessels. Time on commercial ocean vessels is money, in term of salaries, fuel maintenance, etc.

CHAPTER 17

MAINTENANCE OF METEOROLOGICAL EQUIPMENT

Accurate environmental observations are the basis for accurate forecasts. Without properly functioning observational equipment for gathering environmental data the accuracy of the observation is invariably going to become questionable. This means that the instruments involved must be maintained in the best possible operating condition at all times.

Aerographer's Mates first class and chief are expected to perform operator's tests, calibrations and adjustments to meteorological equipment excluding electronic components. They are also expected to inspect work areas for potentially hazardous conditions and practices; and interpret directives and instructions on safety precautions applicable to work areas and equipment within their area of responsibility.

This chapter will be confined to information pertaining to the above stated qualifications. Information pertaining to description, purpose, and operating instructions for meteorological equipments may be found in AG 3 & 2, NavTra 10363-D or the appropriate instrument technical manuals.

MAINTENANCE PROCEDURES

At the present time a few meteorological equipments are maintained under the Navy Maintenance and Material Management System (3M System). Information pertaining to the 3M System may be found in the latest revision to PO 3 & 2 and PO 1 & C training manuals or by reference to the Standard Navy Maintenance and Material Management System Manual, OpNav 43P2.

LOCAL MAINTENANCE PROGRAM

The maintenance support of meteorological equipments is normally the responsibility of the host or local command. Separate NavAir instructions delineate the maintenance responsibility for meteorological equipment installed in aircraft. Operator's preventive maintenance is the responsibility of the Aerographer's Mate.

Preventive maintenance is the systematic care and inspection of all meteorological material for the purpose of retaining it in serviceable condition. By detecting and correcting minor incipient failures before they develop into major defects or malfunctions, the operational service life of any piece of equipment can be extended considerably. Preventive maintenance on electronic equipment must be accomplished by permanently assigned qualified technical personnel who are thoroughly familiar with the operational characteristics of the equipment. The procedure for adequate preventive maintenance checks is described in the Maintenance instruction Manual for each of the complex equipment. Strict adherence to this procedure is necessary for the peak performance of the equipment. Figure 17-1 illustrates a sample daily preventive maintenance schedule.

A modification of this schedule is presented in chapter 18. Similar schedules for weekly, semi-monthly, monthly, quarterly and annual checks should be prepared on each piece of appropriate equipment.

A maintenance log should be maintained for each piece of meteorological equipment, especially items of complex equipment. Time and date

DAILY PREVENTIVE MAINTENANCE SCHEDULE

	*1	2	3	4	5	6	7	*8	9	10	11	12	13	14	*15	16	17	18	19	20	21	*22	23	24	25	26	27	28	*29	30	31
MERCURIAL BAROMETER																															
1. CLEAN CASE AND EXTERNAL SURFACES	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM	HQM
RADAR SET																															
1. CLEAN ALL SURFACES OF SET	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT
2. CLEAN ALL SCOPES WITH WINDEX	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT	CAT
AN/UMQ-5																															
1. RECORDER CHART OPERATION	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO	JCO
2. INKING SYSTEM	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB	HEB
3. VANE ALINEMENT	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ	JHZ
4. AGREEMENT OF ALL RECORDERS & INDICATOR	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm	Jm

*CHECK WEEKLY MAINTENANCE SCHEDULE

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Figure 17-1.—Sample preventive maintenance schedule.

of all emergency, routine, and/or preventive maintenance should be logged along with the parts replaced. Cause of outages and the remedy therefore should be noted.

METEOROLOGICAL AND OCEANOGRAPHIC EQUIPMENT PROGRAM (MOEP)

The Meteorological and Oceanographic Equipment Program (MOEP) was established to insure that meteorological and oceanographic equipment is installed, operated, and maintained in the most effective, reliable, accurate, and economical manner. The program is staffed by specially trained,experienced officers designated as Meteorological and Oceanographic Equipment Technical Liaison Officers (MOETLO's). Additional personnel include a select group of civilian engineers and military technicians. These personnel, military and civilian, provide fleet and shore-based commands with the specialized assistance required in connection with meteorological equipment.

The administration, organization, and areas of responsibility are outlined in the U.S. Navy Meteorological and Oceanographic Support Manual, NavWeaServComInst 3140.1 ( ).

All ships, stations, and Marine Corps Aviation Field Activities concerned with the installation, operation, maintenance and rework of assigned meteorological and oceanographic equipment are expected to utilize the services of MOEP personnel. The service provided will be in the form of assistance in solving meteorological and oceanographic equipment problems beyond the capability of local maintenance personnel.

The ship or station should arrange for local maintenance personnel to be present for on-the job training and to assist MOEP personnel during their visits. Whenever possible they should utilize the informal shop-classroom technician training offered at MOEP activities as listed in the support manual, NavWeaServComInst 3140.1( ).

The ship or station will also be required to maintain maintenance records on assigned meteorological and oceanographic equipment.

Requests for Assistance

Commanding officers are requested to indicate their requirements for planning, installation, and/or technical assistance whenever the need arises to the appropriate activity. All requests must include the following information:

1. Name and exact location of the activity to which the engineer is to report (for ships, inclusive dates of availability at a specific location).
2. Type and model of equipment on which assistance is required.
3. Nature of technical difficulty and availability of spare parts if required.

Spare Parts

The NavSup Manual charges supply officers with the responsibility for stocking spare parts listed as allowance items in the NavAir 00-35QL-40/50/60 Initial Outfitting List series. Parts of a consumable nature listed in NavAir Allowance List 00-35QL-22 are still issued directly to meteorological units, upon receipt, by the supply officer. Meteorological units afloat and ashore are urged to avail themselves of this concept by initiating action to have spare parts stocked in local supply to the extent allowed by the allowance list.

MAINTENANCE MANUALS

Throughout the next three chapters, reference is made to consult the appropriate technical manual for more complete details on operation, service, and maintenance of the equipment mentioned. Such manuals will be listed in one of two places—the Navy Stock List of Forms and Publications, Cognizance Symbol I, Section VIII, Part C, NavSup Publication 2002, or in NavAir 00-500A, Naval Aeronautic Publications Index, Equipment Applicability List.

NavSup 2002, SEction VIII, Part C, contains a complete numerical listing of all available naval aeronautic manuals distributed by NavSup and stocked for issue as of the date of publication. All manuals are listed by both code number and title.

These publications stock lists are supplemented bimonthly. Each supplement contains a listing of publications distributed or canceled since the issue date of the stock lists.

In NavAir 00-500A, all maintenance manuals for meteorological equipment are listed in alphanumerical order under the heading "Meteorology Instruments."

Publications Numbering System

Coded numbers are assigned to publications issued by NavAir to orderly divide this material into proper subject categories. Knowledge of these subheadings will more readily enable the user to determine the desired information in the minimum amount of time. A brief explanation of the publications numbering system is given in the following paragraphs.

For manual type publications the number consists of a prefix and a series of three parts. The prefix consists of letters which identify aeronautic publications under the cognizance of NavAir.

The prefix may consist of any of the following:

1. NavAir—This prefix is assigned to those technical publications originated by the Naval Air Systems Command. In the stock list, it is shortened to NA.
2. NavAer—This is the prefix assigned to the old Bureau of Aeronautics manuals that are still in effect. To conserve space in stock lists it has been shortened to NA.
3. NavWeps—This prefix is assigned to technical publications originated by the now abolished Bureau of Naval Weapons. It is abbreviated NW.
4. AN—This prefix was previously assigned to manuals used by the Navy and the Air Force when such were prepared to coordinated military specifications. It is no longer used on new publications, but the ones in issue remain in effect.
5. TO—This is the prefix assigned to an Air Force originated technical publication.

6. CO—This was previously used to designate a technical publication with a Confidential security-classification. It is no longer assigned to new material, but remains in effect for existing publications until superseded.

7. TM—This is the prefix assigned to an Army originated technical manual.

The three parts which make up the remaining portion of the number are utilized to indicate the following:

Part I. This consists of numbers to identify the general subject series with the basic subject to which they pertain.

Part II. This consists of numbers and/or numbers and letters to designate the specific class, group, type, or model and manufacturer of the equipment. The subject breakdowns for this part are found listed in the beginning of each separate major division within the stock list.

Part III. This consists of numbers to designate a specific publication. For airframes and engines this part designates a specific type of publication. For other types of equipment this part is assigned in numerical sequence and has no reference to the type of publication.

Meteorological publications are found in the 50 series and meteorological instruments generally in the 50-30 series, with some in the 16-30, 16-35, and 16-45 series.

Operation and Service Instructions Manual

This manual gives the operation and service instructions for the particular item covered. Also included are instructions for installation, service, inspection, maintenance, and lubrication.

Overhaul Instructions Manual

This manual contains instructions for complete overhaul of the instrument or equipment as performed by the major repair facilities.

Illustrated Parts Breakdown

This is actually a parts catalog which lists all of the parts of the complete equipment. It is designed to enable maintenance personnel to identify and order replacement parts for the model of equipment covered by the publication. It contains a GROUP ASSEMBLY PARTS LIST and a NUMERICAL PARTS LIST.

To find a part when the name of the part is known, turn to the table of contents, find the name of the part, and turn to the page indicated.

Meteorological Equipment Maintenance Manuals

These manuals are currently being written by the Naval Aviation Engineering Service Unit (NAESU) for certain meteorological electronic and electro-mechanical equipments in use or under procurement by the Naval Air Systems Command. They are written for the AT/ET2 level and cover intermediate type maintenance procedures. (Intermediate type maintenance is component repair maintenance.) Several have been written and issued at the time of this writing. Simple operator maintenance procedures for AG personnel are also included.

TEMPERATURE INSTRUMENTS

Thermometers are not repairable items, but require certain maintenance services to maintain accuracy and ease of reading. The most effective maintenance is careful handling. Thermometers and psychrometers are delicate, precision instruments and must be protected from shocks and jars.

VISUAL CHECKS

Visual checks are made to note any defects in the thermometer or in the metal back upon which it is mounted. Be especially careful to note at the time of each observation, whether any dirt, moisture, or any other foreign matter is present on the thermometer bulb or tube; and whether any separations exist in the fluid column.

REPLACEMENT

If a thermometer is found to be inaccurate, it must be replaced with one that is known to be accurate.

For care and maintenance of the thermometers, see NA 50-30FR-518, Operation and Maintenance Instructions.

TOWNSEND SUPPORT (ML-54)

Lubrication

Once each month, place one or two drops of MIL-O-6085 oil in the oilhole in the maximum thermometer holder. Do not overlubricate.

HUMIDITY INSTRUMENTS

The humidity instruments discussed in this section are the sling and electric psychrometers. They are delicate, precision instruments and should be treated as such.

CARE OF PSYCHROMETERS

The wick on the psychrometer wet bulb should be replaced at least once a month, and more often when local atmospheric conditions cause a rapid collection of dirt and foreign matter on the wick. Psychrometers used on board ship collect salt very rapidly, and daily replacement of the wick may be necessary to obtain accurate readings.

The metal back upon which the thermometers are mounted should be cleaned at least once every 3 months, or more often if necessary. Excess lubricant, dirt, and other foreign matter should be removed from the psychrometer sling assembly at least once a month.

Once each month, place one or two drops of MIL-O-6085 oil on the sling psychrometer swivel link. Do not overlubricate.

ELECTRIC PSYCHROMETER ML-450A/UM

Although the electric psychrometer is constructed primarily of noncorrodible materials, prolonged exposure to weathering, salt air, stack gases, and other corrosive elements will shorten the useful life of the instrument. The instrument should, therefore, be sheltered when not in actual use. Whenever the electric psychrometer is to be stored or used infrequently the batteries should be removed to prevent corrosion.

INSPECTION

Inspect for cracks or breaks in the plastic parts. Inspect threaded holes and screws for wear. Inspect the fan for damage. Inspect the thermometers for cracks and separated mercury columns. Inspect the carrying case for damage. Inspect the motor shaft for smoothness of rotation. For troubles, probable causes, and remedies, refer to the technical manual.

Repairs or Replacement

All plastic parts that are broken, cracked, or have missing or damaged threaded inserts should be replaced. Damaged screws should be replaced. If the motor shaft does not turn freely or if the fan is damaged, the motor and fan assembly should be replaced. If the contacts on the contact block do not contact the motor terminals correctly, they should be carefully bent so that good electrical contact is insured. If one thermometer is broken, both thermometers must be replaced with a new matched set (two thermometers which read the same without wicking attached). A carrying case that is damaged beyond use should be replaced.

TESTS

To test the instrument, first install fresh battery cells. Then turn the control knob on the rheostat-switch in a clockwise direction. The motor should run, causing the fan to draw air into the sliding air intake and force it out of the exhaust ports. If this does not happen, the battery cells are improperly installed. Proper installation of the battery cells should correct the trouble.

As the control knob is turned in the clockwise direction, illumination should increase. If it does not, check the rheostat-switch and lamp. If either is defective, replace it.

PRESSURE INSTRUMENTS

MARINE BAROGRAPH

Maintenance

Use a damp cloth to clean the plastic sheet window in the case. Do not use a solvent cleaner or a dry cloth, as either can damage the plastic pane. Very little other maintenance is required. Under normal operating conditions this instrument should be cleaned well once a year. The element cover should be removed only to clean out any bulky dirt, cobwebs, etc. Do not wipe out this mechanism. Check the pen for wear or roughness and replace as necessary.

The chart drive mechanism should be cleaned and oiled annually. This oiling should not be attempted by personnel unless they are properly instructed in doing it.

For other troubles and remedies with this piece of equipment, consult the appropriate NavAir technical manual.

FORTIN BAROMETER

Maintenance

Only minor repairs should ever be attempted by meteorological personnel. Mercury leakage, broken tubes, damaged instruments, or other faults that affect the accuracy of the readings are cause for immediate survey and replacement of the instrument.

The following minor repairs may be made locally as long as no damage has occurred that will otherwise affect the accuracy of the instrument.

BROKEN THERMOMETER.—Remove the plate that holds the attached thermometer. Take out the broken stem by removing the two small screws in the bracket at the top of the stem. Insert a new thermometer with the bulb projecting inside the barometer case. Replace the screws and the plate.

BROKEN CYLINDRICAL GUARD GLASS.—The guard glass may be replaced unless the scales have been displaced or injured. Remove the four screws holding the top fitting. Remove broken glass and carefully replace, taking care not to disturb scale settings. Any slight jar might displace the scale settings and introduce a constant error in readings. Replace the top fitting, making sure the cork support for the top of the tube is correctly placed.

CLEANING.—The cleaning of this instrument is restricted to wiping the case and external surfaces of the barometer regularly with a clean, lint-free cloth, lightly dampened with water. Occasionally, the scales may be wiped clean and a thin coat of high grade clock oil applied.

Preparation for Reshipment

When a Fortin barometer is to be moved over a considerable distance, it is necessary to transport the instrument in an inclined or inverted position, cistern uppermost.

An adjusting screw is provided beneath the barometer cistern for the purpose of raising the mercury during the process of inverting the barometer for shipment. The procedure to be followed during the inverting process is of critical importance and must be followed exactly to avoid damage to either the leather bag in the cistern or the glass tube. Refer to the technical manual for specific instructions and precautions to be followed. After the barometer has been inverted, with cistern uppermost, the cistern screw should be loosened about one or one and one-half turns in order that there may be sufficient free space for expansion of the mercury in the event of an increase of temperature. The barometer may be safely transported or carried by hand in an inverted or inclined position, as long as care is used to avoid subjecting it to rough handling.

The inside of the barometer case should be well packed with excelsior or other packing material and the cover securely closed. The case is then wrapped well with heavy wrapping paper.

A special wooden packing case is used for shipping the inverted barometer and its case. The packing case must be carefully packed with excelsior or other packing material, insuring that the box is completely filled. The cover is then securely fastened with screws. Cross braces should be attached at the bottom, projecting at least a foot from the sides to insure shipment with the barometer remaining in a vertical position. The case should be labeled plainly in red paint or red printed letters: HANDLE WITH CARE, DELICATE INSTRUMENTS, GLASS, VERY FRAGILE, and THIS END UP. Labels should be in large letters on all four sides of the box.

Tests

The accuracy of the mercurial barometer is tested by comparison with a pressure standard of greater accuracy. Procedure for this comparison may be found in the Manual of Barometry (WBAN), NW 50-1D-510.

PRECISION ANEROID BAROMETER (ML-448/UM)

The Precision Aneroid Barometer, ML-448/UM, is used in the Semi-Automatic Meteorological Station AN/GMQ-14( ) and aboard ship.

Checks and Adjustments

Prior to its initial use in a weather office, a comparison check with an approved standard barometer is required in accordance with the procedure presented in FMH No. 1, Surface Observations. After a correction value has been obtained, periodic checks are made to ascertain the instrument's reliability.

NOTE: The only authorized adjustment to be made in the field on this instrument is the current pressure adjustment.

Maintenance

The exterior of the case should be dusted whenever required and the dial window should be wiped with a clean damp cloth if necessary. Inspect the general physical appearance of the instrument. A cracked dial window, dents, bends, and other external physical damage probably indicate a need for overhaul of the instruments since impact sufficient to cause external damage is usually sufficient to render the instrument inoperative or of suspect accuracy.

No repair or parts replacement is to be attempted at the observer's maintenance level. No attempt to lubricate the instrument is to be made at this maintenance level.

Preparation for Reshipment

If necessary to package the precision aneroid for shipment, it should be wrapped in heavy paper, closed with tape, overwrapped with at least 2 inches of cushioning material, and then placed in a snug-fitting corrugated cardboard box. For further protection in shipment, the packaged barometer should be placed in a wooden shipping crate and protected with additional cushioning material.

Tests

The barometer should be checked by carefully raising and lowering it through a distance of 8 to 10 feet without jarring. If the pointer does not indicate a preceptible change due to this change in height, the barometer requires servicing by an instrument shop and should be returned for repairs.

The accuracy of the instrument is determined during the comparison noted in the section entitled "Checks and Adjustment." Failure of the barometer to track with the approved standard barometer indicates a lack of accuracy and the need for recalibration or repairs in an overhaul shop.

RECONNAISSANCE ANEROID BAROMETER (FA-112)

The Reconnaissance Aneroid Barometer (FA-112) is briefly described in chapter 8 of the latest revision to the AG 3 & 2 Rate Training Manual. A more detailed description may be obtained from the instruction booklet provided by the manufacturer.

Maintenance

Since the dial of each barometer is individually calibrated for its particular mechanism, very little maintenance can be done in the field. Scale adjustment is necessary when the barometer must be reset to agree with a known standard of accuracy. A small bladed screw driver inserted in the slot of the adjusting disk, which is flush with the dial, is used for this prupose. A counterclockwise rotation of the disk causes a like movement of the pointer. The adjusting disk is accessible thru the screw plug in the transparent dial cover of the instrument.

Glass dial covers may be cleaned by wiping with a damp cloth.

The barometer has a filter plug which contains a stainless steel filter screen. This screen should be cleaned occasionally to remove any foreign material. The plug may be removed with a screwdriver. If the orifice which admits outside pressure to the case is found to be plugged, it should be opened by the use of a small diameter wire. The screen should then be cleaned with gasoline, after which it should be moistened with oil and drained before replacing in the barometer case.

Since these instruments are calibrated individually, field repairs are not recommended. In emergencies, certain operations which are outlined in the instrument handbook are possible. These include removal of the dial cover, pointer, and the instrument from the case; and readjustment of the rack for tooth engagement.

The FA-112 barometer has been constructed just as ruggedly as the service for which it is intended will permit. However, in an instrument as sensitive and accurate as this barometer, delicate parts must be used. This barometer must be handled with care.

Lubrication

THE MECHANISM DOES NOT REQUIRE OIL. Oil will only interfere with proper functioning and introduce serious errors in readings.

THEODOLITES

SHORE TYPE

The shore type theodolite is illustrated in figure 17-2 for ease in identifying its various component parts.

Maintenance

There are very few wearing parts on the theodolite. If the instrument is properly handled, carefully packed in its carrying case when not in use (fig. 17-3), covered with the canvas hood for protection when out of the case, and kept wiped clean of lint and dust, there should be very little need for overhaul or replacement of parts, although periodic adjustments may be required.

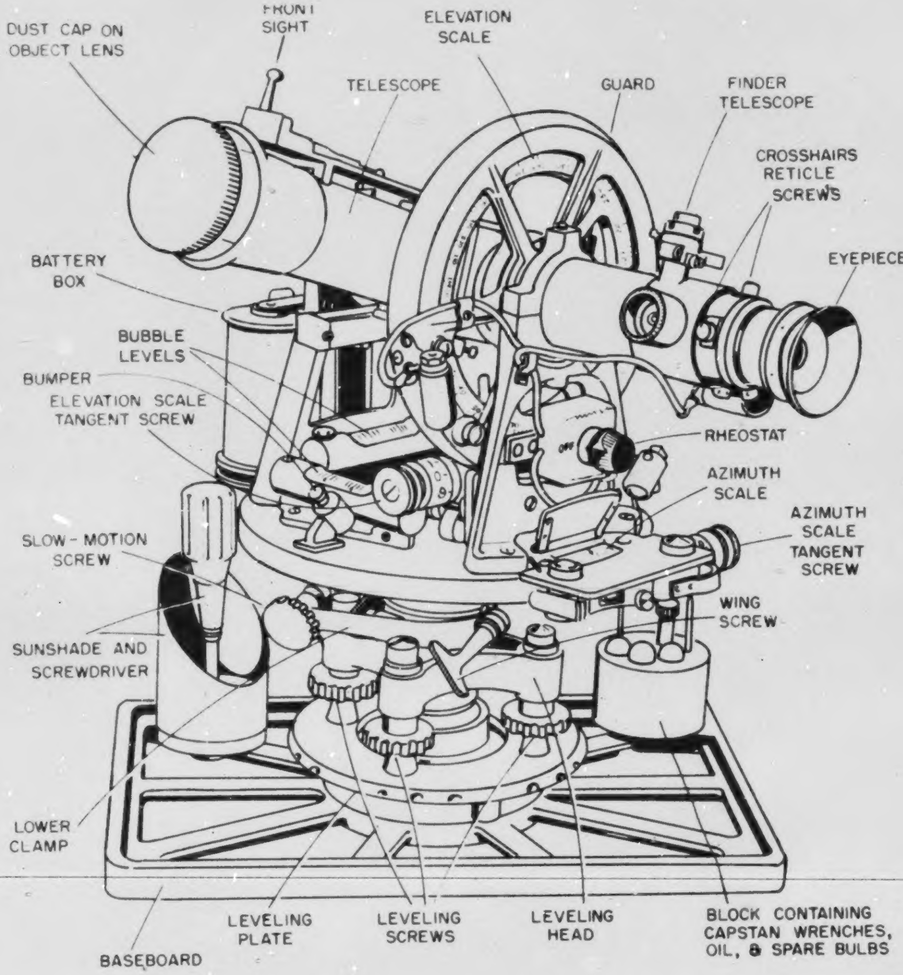
To clean the telescope lenses, first use a clean, soft-haired brush to lightly brush off the lens surfaces. Then wipe clean with a special tissue provided for the purpose. If special tissue is not available, use clean, dry chamois or soft toilet tissue. It is very easy to ruin a lens surface by scratching; therefore, be especially careful when wiping it. Do not wipe hard.

When inspecting the object lens, do not attempt to remove the object lens from its mounting. If the retaining screw has loosened so that the lens is not tight in its mount, the screw may be tightened by means of the Spanner wrench used for this purpose. Handle the object lens and barrel very carefully and inspect the fine screw threads to see that they were not damaged by being crossed when the assembly was installed.

Do not attempt to disassemble the eyepiece assembly, as any disassembly or attempt at disassembly may disarrange the lenses used to make up the eyepiece and render the assembly useless unless it is properly adjusted with precise optical equipment. Examine the assembly carefully for damage and see that the eyepiece moves freely in the spiral groove in the eyepiece adapter.

The silver surfaces of the vertical and horizontal circles may become tarnished from too frequent contact with the hands. Touching the scale surfaces leaves a deposit of moisture and oil that tends to oxidize the surface. The surface can be brightened to some extent by rubbing with boneblack or by applying a few drops of MIL-L 7870 oil and leaving it on the scale overnight. Then wipe the scale clean with a clean, soft cloth. Leave a thin film of oil on the surface to help keep it bright. The horizontal circle scale surface is covered and does not require particular attention. Never use prepared commercial polishes in any form on the graduated scale surfaces of the theodolite.

When cleaning the tangent screw mechanisms, particular care must be taken to remove all dust and grit from the worm gears that move the horizontal and vertical circles. Use a soft cloth or a toothbrush moistened with solvent, such as alcohol, for this purpose.


DUST CAP ON OBJECT LENS
FRONT SIGHT
ELEVATION SCALE
TELESCOPE
GUARD
FINDER TELESCOPE
CROSSHAIRS RETICLE SCREWS
EYEPIECE
BATTERY BOX
BUBBLE LEVELS
BUMPER
ELEVATION SCALE TANGENT SCREW
RHEOSTAT
AZIMUTH SCALE
AZIMUTH SCALE TANGENT SCREW
SLOW-MOTION SCREW
WING SCREW
SUNSHADE AND SCREWDRIVER
LOWER CLAMP
LEVELING PLATE
LEVELING SCREWS
LEVELING HEAD
BLOCK CONTAINING CAPSTAN WRENCHES, OIL, & SPARE BULBS
BASEBOARD

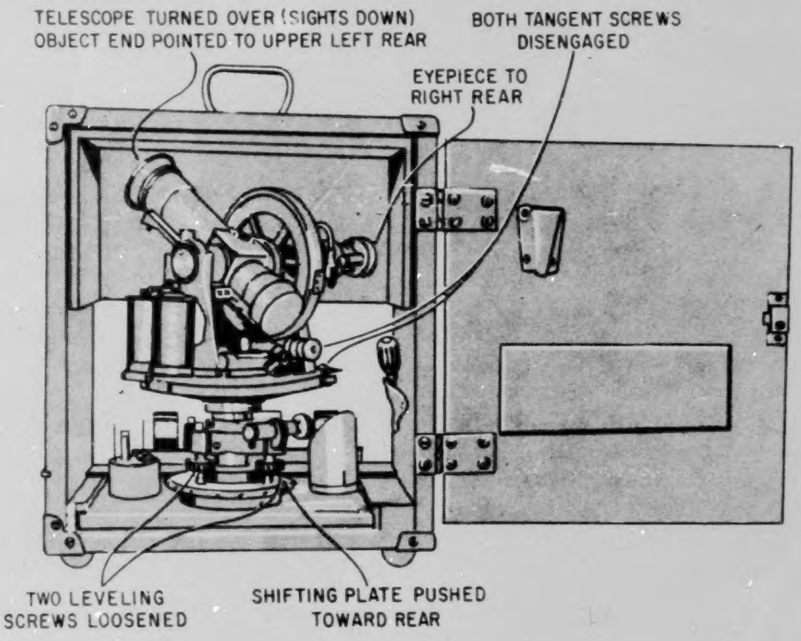
AG.732

Figure 17-2.—Shore type theodolite and baseboard assembly.

Tests and Adjustments

The theodolite is carefully tested and adjusted before issue. However, rough handling or usage may result in the instrument getting out of adjustment. All theodolites should be given a "General Test" as outlined in FMH No. 5, Winds Aloft Observations, upon receipt and quarterly thereafter.

Other tests and adjustments include the spirit level check and adjustment, collimation adjustment, adjustment of horizontal axis, adjustment of vertical circle fiducial mark, and adjustment of tangent screw verniers. Most of these are precision adjustments and require skill, patience, and experience on the part of the person making the adjustment. Extreme care must be taken not to damage the theodolite, and adjustments should be made by a qualified instrument man only. Consult the appropriate technical manual for a more complete discussion of the tests, checks, and adjustments.


TELESCOPE TURNED OVER (SIGHTS DOWN)
OBJECT END POINTED TO UPPER LEFT REAR
BOTH TANGENT SCREWS DISENGAGED
EYEPIECE TO RIGHT REAR
TWO LEVELING SCREWS LOOSENED
SHIFTING PLATE PUSHED TOWARD REAR

AG.733

Figure 17-3.—Theodolite in case.

TRIPODS

Maintenance

The tripods require very little maintenance. The corrosion on the metal legs of the shipboard tripod can be removed with silver polish.

Periodically apply a small amount of oil to the movable joints on the tripods. Remove any excess oil to prevent dust and dirt from collecting on these joints.

CLINOMETERS

ML-119 (SHORE TYPE)

Maintenance

When the clinometer is not is use, keep it in its case to protect the instrument from dust and dirt. Keep the pendant clamped to prevent it from being damaged by sudden movements or jolts. Remove dust from the instrument with a clean soft cloth. If the cover glass on the clinometer is fingerprinted or dirty, wipe the outside with a damp cloth and polish with a soft cloth or tissue.

TESTS

Check the accuracy of the clinometer every month as follows: With the sighting tube in a horizontal position, rest the front edge of the quadrant scale plate on a level surface and check the instrument. The pendant should indicate 0°. If the instrument does not register accurately, return it to the repair depot for adjustment.

ML-591/U (SHIPBOARD TYPE)

Maintenance

Maintenance of the Clinometer ML-591/U which is designed for shipboard use consists primarily of checking the accuracy and performance periodically on a 30 day cycle. There are specific steps to be followed when accomplishing the performance test. The proper procedure is described in the technical manual.

The ML-591/U should also be checked periodically for damage, wear, and corrosion. Maintenance is limited to cleaning, minor repair, replacement of damaged or worn parts, and adjustment. Instructions for major repairs and disassembly are contained in the technical manual.

WIND EQUIPMENT

WIND MEASURING SET AN/UMQ-5( )

Under normal conditions Wind Measuring Set AN/UMQ-5( ) will require very little testing or adjusting. In this section, only the maintenance that Aerographer's Mates First Class and Chief normally perform is discussed.

Maintenance

Maintenence is limited to service, replacement, adjustment, and minor repair which can be performed without disassembly or with partial disassembly not requiring the use of overhaul facilities. Instructions for major repair are contained in the overhaul instruction manual. Such work is to be performed in overhaul shops only.

COMPLETE WIND MEASURING SET AN/UMQ-5( ).–Service inspection of the complete set consists of a daily observation of the detector and indicator behavior and a monthly check on cables, connections, and mechanical security.

The daily inspection consists of a check to see that the vane alines the detector in the apparent wind and that the impeller turns freely. Compare the indications of all recorders and indicators on the detector circuit. They should be in agreement.

The monthly check consists of checking the cables for physical damage or deterioration. Check the mechanical security of component mountings.

When damage to cabling is discovered during service inspection or indicated during troubleshooting, the cable should be checked and repaired or replaced as necessary. Particular care must be taken to avoid improper connections, since errors introduced to the indicators/recorders as a result of improper connections are not always readily apparent.

DETECTOR.–There will be occasions when the detector direction synchro will need zeroing; that is, when the vane points to north, the indicators and recorders should read north. The procedure for zeroing the components of the UMQ-5( ) may be obtained from the instrument technical manual.

Repainting by Aerographer's Mates is restricted to minor touchup of scratches on the detector vane and housing and complete refinishing as necessary on the support. Care must be taken not to apply too much paint anywhere on the detector vane since it may unbalance the instrument and cause erroneous indications.

INDICATORS.–The only service inspection required for the indicators is an observation of their operation to detect erroneous readings, or failure of lamps.

No special procedures are required for repair and replacement of parts in the indicators.

RECORDER.–Service inspections on the wind direction and speed recorder consist of scheduled checks and inspections. Table 17-1 lists the scheduled times and the nature of the inspection.

For a further discussion on recorder maintenance, see the Handbook of Operation and Maintenance Instructions, NA 50-30FR-525.

Component Tests

Most of the electrical tests on the Wind Measuring Set AN/UMQ-5( ) are performed by the ship or station electronics personnel. Tests which can normally be performed by Aerographer's Mates are presented in this section.

COMPLETE SET.–For the wind direction circuit test, position the wind direction-speed detectors so that the vane is pointing the nose directly toward north. This may be accomplished by clamping a straightedge to the machined side of the connector housing and alining the vane with the straightedge. All indicator direction pointers and recorder direction pens in the same circuit with the detector should indicate north.

Table 17-1.—Recorder inspection schedule.

Item	Nature of inspection	Time
Recorder chart	Inspect for clear, legible record trace, proper time setting, sufficient chart reserve, takeup without binding, and agreement of recorded values with indicator readings.	Daily.
Pens	Check for evidence of clogged ink, fuzzy line on record, recording on sudden swings.	Daily.
Ink tanks	Check to see that sufficient ink is in tanks.	Weekly.
Complete instrument	Check for evidence of dust, dirt, or corrosion.	Monthly.
	Check operation in accordance with the section entitled "Component Tests."	Quarterly.
	Check lubrication in accordance with the applicable BuWeps technical manual.	Quarterly.

DETECTOR.—Only tests of a mechanical nature which are performed by Aerographer's Mates are discussed in this section. As stated earlier, electrical tests are normally performed by the ship or station electronics personnel.

When checking the wind direction-speed indicator for excess friction, remove the detector from the connector housing. Attach the protecting cover to the top of the connector housing to prevent foreign matter from entering the female connector. Carry the detector inside, away from any effect of the wind.

Attach a test disk (the weight of this disk is equal to the weight of a 1-cent coin) at the tip of one of the impeller blades with a piece of scotch tape. Turn the blade so that it is 45° from the top center. Then release the impeller. The weight of the test disk should be sufficient to carry the blade to the bottom if excessive friction is not present. If the weight of the disk is insufficient to carry the blade to the bottom, it is possible that either the magneto needs replacing, or the impeller holder may be rubbing on another surface.

To check the vane for excessive friction, attach a test disk (the weight of this disk is equal to a 50-cent coin) to the flat side of the tail surface nearest the trailing edge. Hold the vane approximately in line with the bench top; then release. The weight of the disk should be sufficient to carry the tail surface down if excessive friction is not present. If the weight of the disk is insufficient to carry the vane downward it is possible that the synchro, ball bearings, or brushes need replacing. Before making the above vane check, make certain that the vane is in static balance.

INDICATORS.—Tests on the indicators are of an electrical nature. Since these tests are normally performed by maintenance personnel, they are not discussed.

RECORDER.—For the tests on the wind-direction and speed recorder, only the chart drive mechanism is discussed. The chart drive mechanism is tested during operation. If the chart feeds smoothly, takes up without binding, and does not gain or lose too much time, it may be considered to be operating properly.

WIND MEASURING SET AN/PMQ-3( )

Inspection

The only service inspection required is to make certain that the turbine and vane are free to rotate and that there is pointer movement when the turbine is rotated. A method to check for freeness is to hold the instrument in its operating position and walk at a moderate rate in an area where there is no air movement. If the vane assumes the correct position and there is a speed indication on both scales, it is probable that the instrument is in a satisfactory condition.

Maintenance

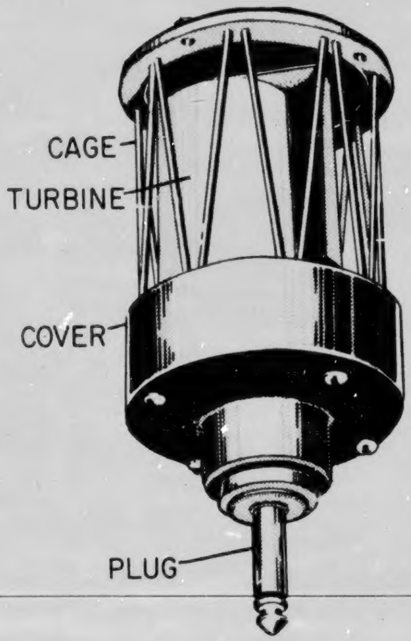
The causes of trouble encountered on the PMQ-3( ) can normally be identified through observance of the characteristics of the wind speed pointer indications. The indicator will present erratic or abnormal wind indications dependent upon which component is defective. The various troubles, the probable causes, and recommended remedies are described in the instrument technical manual.

A trouble can generally be isolated by replacing a component with another known to be in good condition. Table 17-2 includes the components of Wind Measuring Set AN/PMQ-3( ) that the Aerographer's Mates are authorized to install.

Table 17-2.—Replaceable parts list—AN/PMQ-3( ).

Part name	Units per assembly
Wind speed detector	1
Wind vane	1
Sight	1
Speed indicator	1
Trigger and switch assembly.	1
Vane nose	1
Bezel	1
Carrying case	1

WIND SPEED DETECTOR.—To replace the wind speed detector (fig. 17-4), grasp the detector and give it a slight twist in a counterclockwise direction (looking down on the instrument) and pull it off straight. Remove the spare detector from the case in a like manner. Install the spare detector by reversing the above procedure, making certain that the unit is securely locked in position.



AG.734

Figure 17-4.—Wind Speed Detector (AN/PMQ-3( )).

WIND VANE.—Minor defects and dents are not cause for replacement. If twisted parts affect the accuracy, try to straighten them. If the vane is not repairable, the spare wind vane from the case should be installed and another wind vane requisitioned from stock spares.

WIND SPEED INDICATOR.—If physical damage is not visible, and as defective indication may be caused by another component, do not discard the unit until its condition has been proved unsatisfactory by a replacement.

TRIGGER AND SWITCH ASSEMBLY.–If physical damage is not visible, and as a defective indication may be caused by another component, do not discard the unit until its condition has been proved unsatisfactory by a replacement.

The anemometer-wind vane requires no lubrication or cleaning by the operator.

In all cases of replacement or adjustment, consult the current technical manual for complete instructions.

METEOROLOGICAL MEASURING SETS (AN/PMQ-5( ) AND AN/PMQ-7)

For the operating principles and service instructions refer to the Handbook of Operation and Service Instructions, AN/PMQ-5, NA 50-30FR-513, and Handbook of Operation and Service Instructions, Meteorological Measuring Set, NavWeps 50-30 PMQ-7-1.

Aerographer's Mates do not normally come in contact with these equipments; therefore they are not covered in any more detail here.

CEILING LIGHT PROJECTOR ML-121

INSPECTION AND CLEANING

Once a week, and more often if necessary to insure full beam intensity, clean the cover glass on the projector housing and the reflecting surface of the reflectors. Inspect the drainage holes in the mirror and housing, and clean them as often as is necessary to insure adequate drainage and ventilation of the enclosure.

When the sun is shining brightly into the projector, the intensity of heat and light concentrated in the area above the parabolic reflector, especially in the area near its focal point, may be sufficient to burn the skin or seriously injure the eyes. Dark glasses must always be worn when looking into the reflector or at the lighted filament of the lamp.

Liquid glass cleaners or other nonabrasive glass cleaners used with soft clean cloths are recommended for cleaning the cover glass and the reflectors. Avoid scratching or otherwise damaging the reflectors. Replace the cover glass door gasket if water is leaking into the housing.

MAINTENANCE

The maintenance that is performed by Aerographer's Mates will consist of replacing lamps, focusing the lamp, and beam alinement.

WARNING: HIGH VOLTAGE is used in the operation of this and some other meteorological equipment mentioned in other sections of this Rate Training Manual. DEATH ON CONTACT may result if personnel fail to observe safety precautions. Learn the areas containing high voltage in each peice of equipment. Be careful not to contact high-voltage connections when calibrating, adjusting or testing these equipments. Before working inside the equipment, turn power off and ground points of high potential before touching them.

Replacing lamps

Loosen the wing nuts, replace the defective lamps, and secure the access door in accordance with instructions in the instrument technical manual.

Replace defective lamps promptly whenever the lamp has begun to blacken or the filament sags to a noticeable extent. Lamp life will vary considerably due to local conditions and is highly critical with respect to overvoltage or excess voltage fluctuations. The optimum voltage is 11.8 volts and should never exceed 12 volts. Lower voltages decrease the intensity of the spot and the minimum voltage for satisfactory operation is 11.3 volts. The voltage across the lamp should be checked and adjusted if short lamp life or reduced lamp intensity is experienced.

For instructions on focusing the lamp and beam alinement consult the Handbook of Operation and Maintenance Instructions, NA 50-30FR-521.

CHAPTER 18

MAINTENANCE OF AUGMENTING METEOROLOGICAL AND OCEANOGRAPHIC EQUIPMENT

As the scientific fields of meteorology and oceanography continue to advance, so does the importance of rapid and accurate meteorological and oceanographic environmental measurements. Electrical and electronic equipments currently in use continue to be improved as well as augmented by newer and more accurate equipment.

As senior Aerographer's Mates, you must not only be responsible for the correct operation of these equipments but you must also insure that they are kept operating to their fullest extent. An important part of this task entails the periodic performance of operators tests, calibrations, and adjustments. The maintenance of electronic equipments are not performed by Aerographer's Mates. This equipment will be maintained by the ship or station's electronics personnel.

Many of the tests, calibrations, and adjustments to meteorological and oceanographic equipment are discussed in this manual. However, the intent is to provide a general knowledge and not to substitute for the technical manual. Wherever possible reference should be made to the applicable technical manual to insure correct maintenance of the equipment. AG 3 & 2, NavTra 10363-D should be reviewed where necessary for general information relating to description, theory, or principles of operation.

TRANSMISSOMETER SET AN/GMQ-10( )

If accurate measurements of visibility are to be obtained through utilization of the Transmissometer Set AN/GMQ-10( ), the checks and adjustments described in the following paragraphs should be accomplished by the Aerographer's Mate.

CHECKS AND ADJUSTMENTS DURING NORMAL OPERATION

1. BACKGROUND CHECK. This check measures the effect of background illumination on the transmission reading. Turn the background switch to TEST and hold it there. With this switch in the TEST position, a relay in the projector power supply is energized, thus turning off the lamp. Then the reading of the meter and recorder is that due to background.
2. ZERO ADJUSTMENT. If, after stabilization, the meter does not read zero when the zero switch is thrown to TEST position, adjust the zero adjustment potentiometer until zero is indicated.

Turn the range switch to the HIGH position and note the meter reading. If it no longer reads zero, be sure to zero the meter for the range to be used. Switching the range switch should also actuate the recorder range pen near the right-hand edge of the chart.

PERIODIC CHECKS AND INSPECTIONS

Careful routine checks of the equipment by Aerographer's Mates very often prevent failure under conditions when maintenance personnel are not available. Use of checklists will prevent the omission of pertinent checks. Entry of pertinent data and remarks in their proper places on the checklists will assist in the early detection of malfunctioning equipment and improper adjustments or operations. All adjustments should be recorded. Operating personnel are responsible for daily checks, some of the weekly checks, and a few others that are limited to a mechanical nature. All other checks or corrective maintenance should be performed by maintenance personnel. Checks should be made at least as frequently as indicated and more frequently when local conditions require them.

Daily Checks and Inspections

The daily check of the transmissometer is made at the indicator, using the past record as a basis for checking the field units. It is desirable to make the daily check at approximately the same time every day. A time when the visibility is good should be chosen. A form such as that illustrated in figure 18-1 may be used for such inspections, checks and adjustments. It should be pointed out here that forms readily available through normal supply systems may be modified locally to meet the requirements for maintaining accurate logs and records. The entry of initials indicates that the item or unit is functioning properly. Enter in the remarks section of the form any remarks regarding maintenance, adjustments performed, or weather conditions at the time of the check which may be of value to maintenance personnel conducting further maintenance, or to station personnel in evaluating the operation of the transmissometer.

Weekly Checks and Inspections

Certain weekly checks, inspections, and adjustments, if necessary, are performed by AG personnel. As pointed out previously, some of these may be made by the maintenance personnel. Figure 18-2 illustrates a sample weekly preventive maintenance checkoff list for the transmissometer. Although the inspections and checks are made by AG personnel, the maintenance technicians perform the indicated corrective maintenance in cases in which malfunctioning exists. These checks should be made on the same day and at as nearly the same time as possible each week.

Biweekly Checks and Inspections

In addition to the weekly checks and maintenance, the following checks and maintenance are performed biweekly by AG personnel:

1. RECORDER. Change the chart, taking care to date both the old and new charts. Be sure that the chart is properly positioned and feeding properly. Clean the recorder by blowing out the paper fragments and debris from the recorder case.
2. INDICATOR, RECEIVER, AND PROJECTOR. Clean interior and exterior of cabinets.

Monthly Checks and Inspections

The monthly checks and inspections should be made at 4-week intervals, and should be made to coincide with a weekly or biweekly check. The checks made on the indicator, projector, and receiver are of an electronics nature and are performed by maintenance personnel rather than Aerographer's Mates.

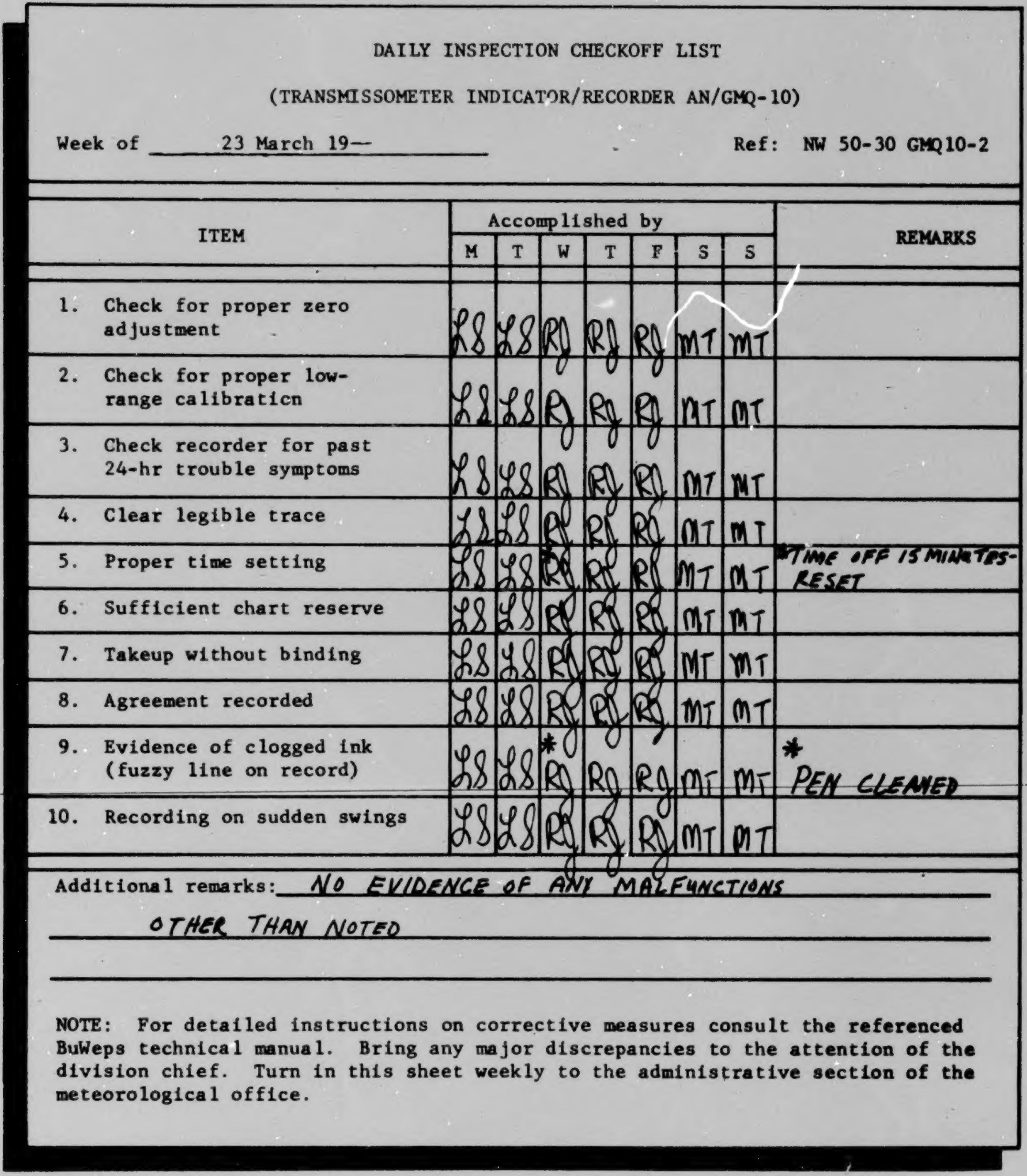
The checks on the recorder will be confined to the writing system. Remove the inkwells and the pen elements from the recorder and wash them with warm water. Pass a stream of water through the pen elements until a clear stream is obtained. Fill the inkwells with the specified type of ink, replace the inkwells, and pen elements, and draw sufficient ink through the pen elements to remove all air bubbles. Check the balance of the pen elements. Check to see that the pen elements do not rub the scale plate or the inkwell at any position throughout its arc.

Quarterly Checks, Inspections, and Preventive Maintenance

The quarterly checks are of an electronics nature and are performed by electronics personnel rather than Aerographer's Mates.

Semiannual and Annual Maintenance

Semiannual and annual maintenance consists of lubricating the transmissometer set and carrying out anticorrosion measures such as cleaning rust from surfaces, touch up, and repainting as necessary.



DAILY INSPECTION CHECKOFF LIST

(TRANSMISSOMETER INDICATOR/RECORDER AN/GMQ-10)

Week of 23 March 19— Ref: NW 50-30 GMQ10-2

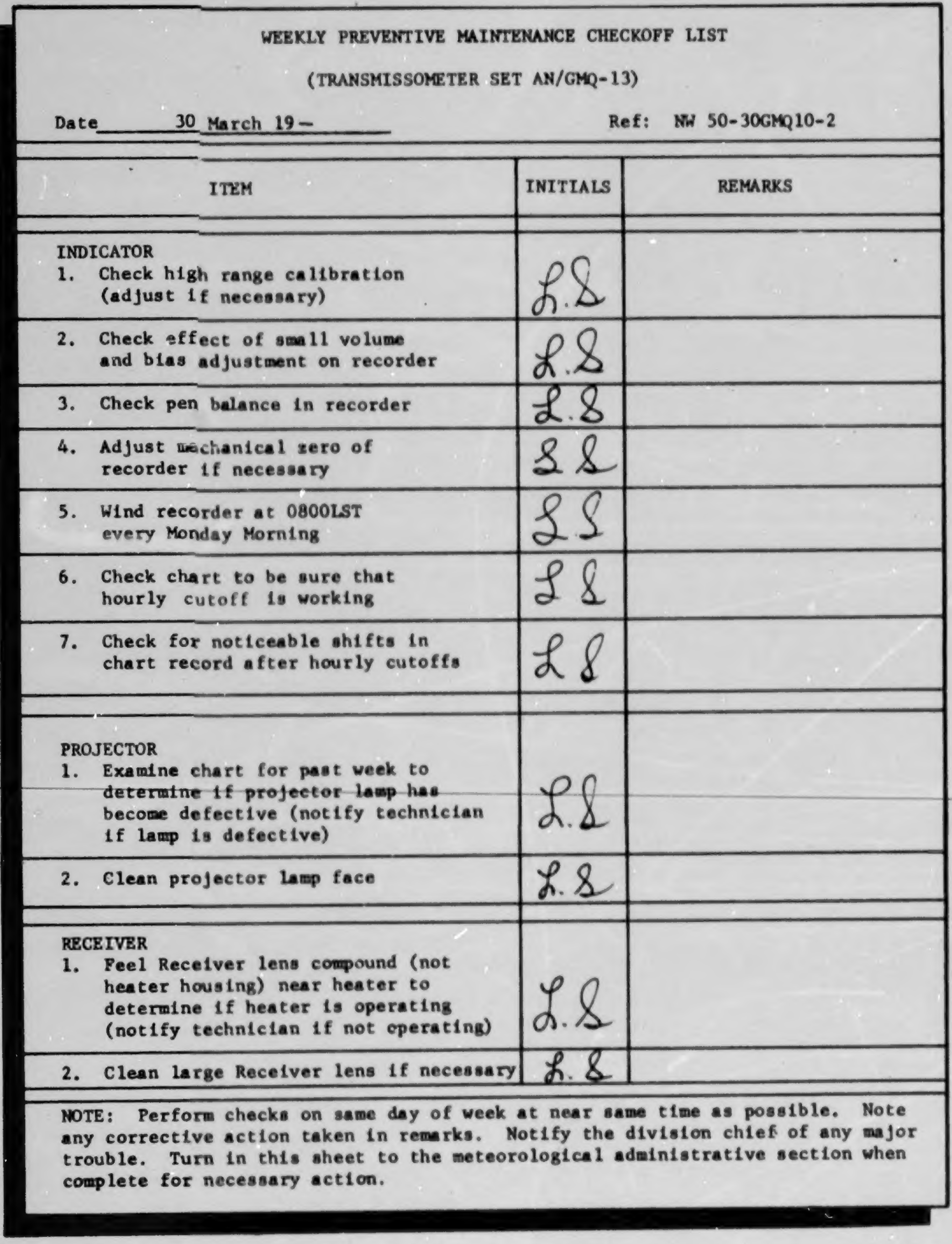
ITEM	Accomplished by							REMARKS
	M	T	W	T	F	S	S	
1. Check for proper zero adjustment	LS	LS	RJ	RJ	RJ	MT	MT	
2. Check for proper low-range calibration	LS	LS	RJ	RJ	RJ	MT	MT	
3. Check recorder for past 24-hr trouble symptoms	LS	LS	RJ	RJ	RJ	MT	MT	
4. Clear legible trace	LS	LS	RJ	RJ	RJ	MT	MT	
5. Proper time setting	LS	LS	*RJ	RJ	RJ	MT	MT	*TIME OFF 15 MINUTES-RESET
6. Sufficient chart reserve	LS	LS	RJ	RJ	RJ	MT	MT	
7. Takeup without binding	LS	LS	RJ	RJ	RJ	MT	MT	
8. Agreement recorded	LS	LS	RJ	RJ	RJ	MT	MT	
9. Evidence of clogged ink (fuzzy line on record)	LS	LS	*RJ	RJ	RJ	MT	MT	* PEN CLEANED
10. Recording on sudden swings	LS	LS	RJ	RJ	RJ	MT	MT	

Additional remarks: NO EVIDENCE OF ANY MALFUNCTIONS OTHER THAN NOTED

NOTE: For detailed instructions on corrective measures consult the referenced BuWeps technical manual. Bring any major discrepancies to the attention of the division chief. Turn in this sheet weekly to the administrative section of the meteorological office.

AG.735

Figure 18-1.—Daily inspection checkoff list for Transmissometer Set AN/GMQ-10 ( ).



WEEKLY PREVENTIVE MAINTENANCE CHECKOFF LIST

(TRANSMISSOMETER SET AN/GMQ-13)

Date 30 March 19— Ref: NW 50-30GMQ10-2

ITEM	INITIALS	REMARKS
INDICATOR 1. Check high range calibration (adjust if necessary)	L.S	
2. Check effect of small volume and bias adjustment on recorder	L.S	
3. Check pen balance in recorder	L.S	
4. Adjust mechanical zero of recorder if necessary	LS	
5. Wind recorder at 0800LST every Monday Morning	LS	
6. Check chart to be sure that hourly cutoff is working	LS	
7. Check for noticeable shifts in chart record after hourly cutoffs	LS	
PROJECTOR 1. Examine chart for past week to determine if projector lamp has become defective (notify technician if lamp is defective)	L.S	
2. Clean projector lamp face	L.S	
RECEIVER 1. Feel Receiver lens compound (not heater housing) near heater to determine if heater is operating (notify technician if not operating)	L.S	
2. Clean large Receiver lens if necessary	L.S	

NOTE: Perform checks on same day of week at near same time as possible. Note any corrective action taken in remarks. Notify the division chief of any major trouble. Turn in this sheet to the meteorological administrative section when complete for necessary action.

AG.736

Figure 18-2.—Weekly preventive maintenance checkoff list for Transmissometer Set AN/GMQ-10( ).

The semiannual and annual lubrication schedules are contained in the Periodic Inspection, Maintenance, and Lubrication section of Technical Manual of Operation and Service Instructions, Transmissometer Set AN/GMQ-10( ) NA50-30GMQ10-2.

To insure that proper maintenance of this equipment is performed at scheduled intervals a job description card, such as the one illustrated in figure 18-3 should be prepared and kept in a tickler file.

TROUBLESHOOTING AND REPAIR

Troubleshooting and repair of Transmissometer Set AN/GMQ-10( ) is limited to those items previously mentioned, and to a few additional items of a mechanical nature. A complete troubleshooting chart may be found in the technical manual for this piece of equipment.

CONVERTER INDICATOR GROUP OA-7900/GMQ-10( )

If the Converter Indicator Group OA-7900/GMQ-10( ) is used to obtain runway visual range (RVR), Aerographer's Mates should insure that minimum acceptable performance standards of operation are adhered to. The theory of operation and operating procedures for this equipment as used in conjunction with the Transmissometer Set AN/GMQ-10( ) are presented in chapter 10, AG 3 & 2, NT 10363-D, and the equipment technical manual.

The following procedure should be followed to determine if the system contains a defective module or circuit:

1. With the power switch turned on, the fan should become operative; the pilot light should light; and when the cover is removed on the encoder, you should be able to observe the encoder disk rotating twice per minute.
2. With the power switch on, the display screen of Digital Display ID-1348/GMQ-10 shown mounted on top of the converter in figure 18-4 should be fully illuminated when the brightness knob is turned fully clockwise.
3. When the "runway light setting" is moved from the "normal" position, the red light indicator on the display panel should light.

The preceding checks may be performed by the operator. However, there are other checks for isolating problems described in the technical manual, which are performed by qualified technicians. In any event, if repairs are deemed necessary, they should be accomplished by the technicians.

The checks mentioned in the preceding paragraphs indicate only that the equipment is operating. To determine if it is operating at the minimum acceptable standards, other checks must be made.

MINIMUM PERFORMANCE TESTS

As already stated, there are many tests which may be performed by qualified technicians. However, there are two preliminary tests which do not take much time and may be performed by the operator. These tests are termed accelerated tests and are referred to individually as the "systems test" and the "display test."

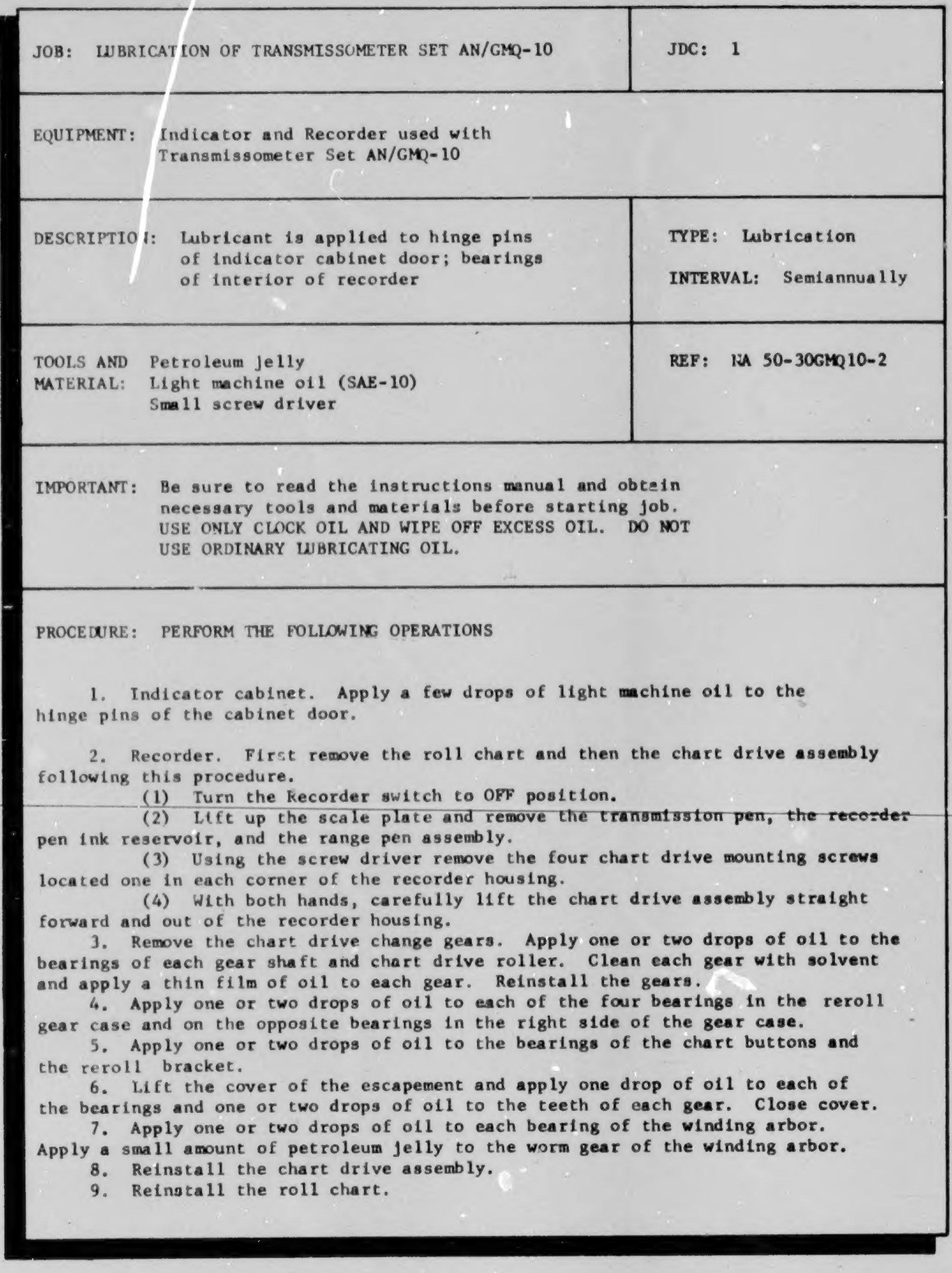
Systems Test

Insure that the converter indicator group (fig. 18-4) is properly connected as described in the technical manual. The transmissometer receiver may be disconnected if desired. Set the converter switches in the various combinations and notice if displayed data is in accordance with table 18-1.

Allow 2 minutes for each display of data and several repetitions to follow in order to observe consistent conversion and display. This test is sufficient to insure that the equipment will operate properly and should be performed at least once a day.

Display Test

The display test is performed in conjunction with the converter as part of the system test. In this test items to be checked are: the readout units, switching relay, and display lamps. As the system is being tested, operation of the readout units can be checked by listening to the clicking of the relays and slide plates, and observing the screen illumination. The clicking, as well as change of readings, should take place once a



JOB: LUBRICATION OF TRANSMISSOMETER SET AN/GMQ-10

JDC: 1

EQUIPMENT: Indicator and Recorder used with Transmissometer Set AN/GMQ-10

DESCRIPTION: Lubricant is applied to hinge pins of indicator cabinet door; bearings of interior of recorder

TYPE: Lubrication

INTERVAL: Semiannually

TOOLS AND MATERIAL:
Petroleum jelly
Light machine oil (SAE-10)
Small screw driver

REF: NA 50-30GMQ10-2

IMPORTANT: Be sure to read the instructions manual and obtain necessary tools and materials before starting job. USE ONLY CLOCK OIL AND WIPE OFF EXCESS OIL. DO NOT USE ORDINARY LUBRICATING OIL.

PROCEDURE: PERFORM THE FOLLOWING OPERATIONS

1. Indicator cabinet. Apply a few drops of light machine oil to the hinge pins of the cabinet door.

2. Recorder. First remove the roll chart and then the chart drive assembly following this procedure.
   (1) Turn the Recorder switch to OFF position.
   (2) Lift up the scale plate and remove the transmission pen, the recorder pen ink reservoir, and the range pen assembly.
   (3) Using the screw driver remove the four chart drive mounting screws located one in each corner of the recorder housing.
   (4) With both hands, carefully lift the chart drive assembly straight forward and out of the recorder housing.
3. Remove the chart drive change gears. Apply one or two drops of oil to the bearings of each gear shaft and chart drive roller. Clean each gear with solvent and apply a thin film of oil to each gear. Reinstall the gears.
4. Apply one or two drops of oil to each of the four bearings in the reroll gear case and on the opposite bearings in the right side of the gear case.
5. Apply one or two drops of oil to the bearings of the chart buttons and the reroll bracket.
6. Lift the cover of the escapement and apply one drop of oil to each of the bearings and one or two drops of oil to the teeth of each gear. Close cover.
7. Apply one or two drops of oil to each bearing of the winding arbor. Apply a small amount of petroleum jelly to the worm gear of the winding arbor.
8. Reinstall the chart drive assembly.
9. Reinstall the roll chart.

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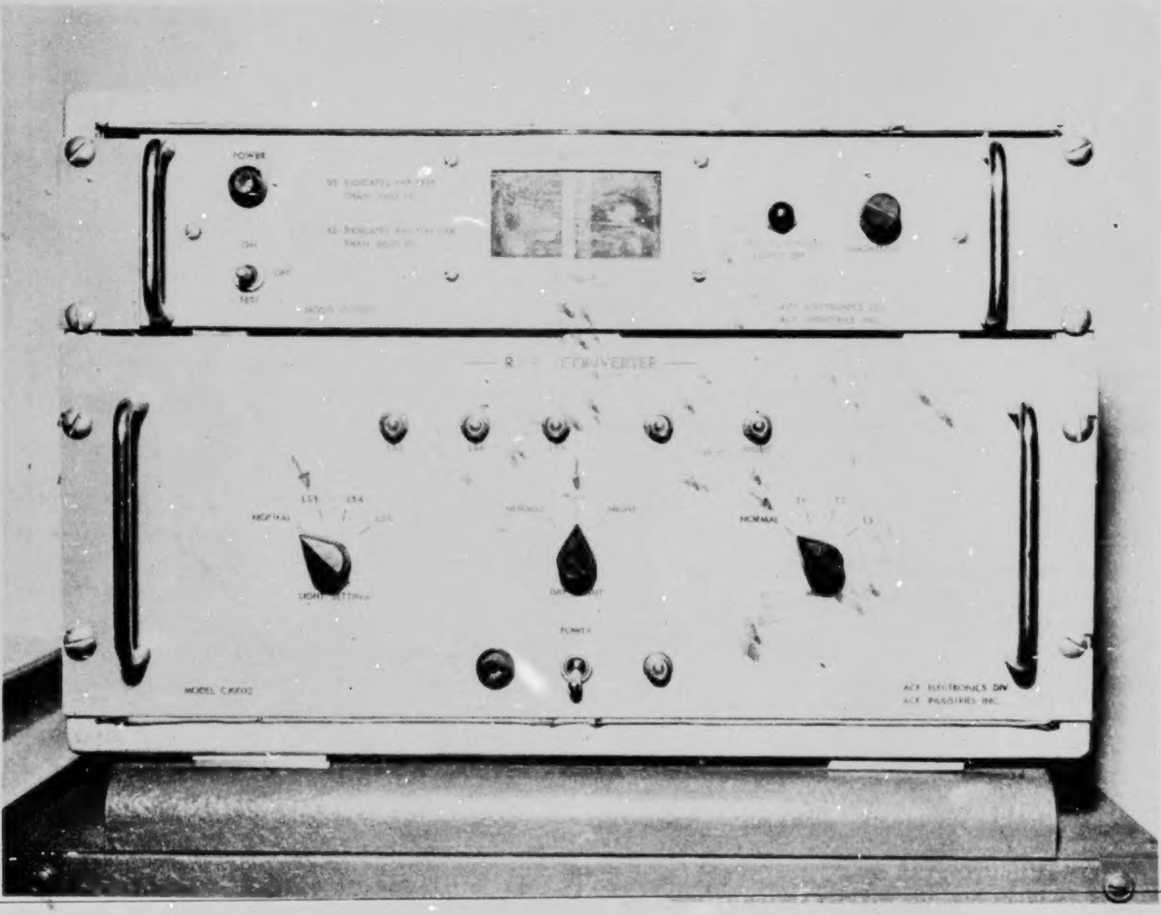
Figure 18-3.–Job description for semiannual lubrication of Transmissometer Set AN/GMQ-10( ).



AG.139

Figure 18-4.—Converter Indicator Group OA-7900/GMQ-10( ).

minute. In case the display lamps are not lighted with the switch in the ON position, flip the switch to LAMP TEST position, and turn the BRIGHTNESS knob clockwise. If they still do not function, notify the technician.

CLOUD HEIGHT SET AN/GMQ-13( )

The AG's responsibility for maintenance, tests and calibration of the Cloud Height Set AN/GMQ-13( ) is very limited. It consists primarily of performing the required weekly checks. This includes insuring that scheduled checks inspections and required preventive maintenance have been performed.

PREVENTIVE MAINTENANCE INSPECTIONS AND CHECKS

To insure that required maintenance is performed on schedule, a preventive maintenance checkoff list should be maintained. This list should follow a similar format to that used for the AN/GMQ-10 as illustrated in figures 18-1 and 18-2 of the chapter. The following paragraphs indicate the checks, inspections, and preventive maintenance which should be performed weekly by the AG's. For details on the accomplishment of these functions, see chapter 10, AG 3 & 2, NT 10363-D, or the Operation and Service Instructions Manual, Cloud Height Set AN/GMQ-13C, NA 50-30GMQ13-3.

Table 18-1.—Switch setting and test data for Converter Indicator Group OA-7900/GMQ-10( ).

Switches			RVR Value Displayed
RVR Light Setting	Day-Night	Mode	
LS3	Day	T1	08
LS3	Day	T2	18
LS3	Day	T3	62
LS4	Day	T1	08
LS4	Day	T2	18
LS4	Day	T3	62
LS5	Day	T1	08
LS5	Day	T2	24
LS5	Day	T3	62
LS3	Night	T1	08
LS3	Night	T2	38
LS3	Night	T3	62
LS4	Night	T1	08
LS4	Night	T2	46
LS4	Night	T3	62
LS5	Night	T1	08
LS5	Night	T2	54
LS5	Night	T3	62

NOTES 1. This table is to be used for the accelerated test only.

2. RVR values displayed may be within +0, - 1 of the values indicated.

3. Switches do not have to be set in the order shown above.

Projector

1. Clean the reflectors and dome cover with a clean soft cloth dampened with ethyl alcohol.
2. Check the lamps for blisters, discoloration, and replace if marred. If lamps need replacing, always replace both and insure 1/8-inch clearance of bulb and shutter; 1/16-inch is minimal.
3. Check for oil leakage near the gearbox and drive motor.

Indicator

1. Clean the air intake.
2. Clean the top plastic with water and a clean cloth.
3. Check calibration weekly to position 2 to check 18-degree markers.

Detector

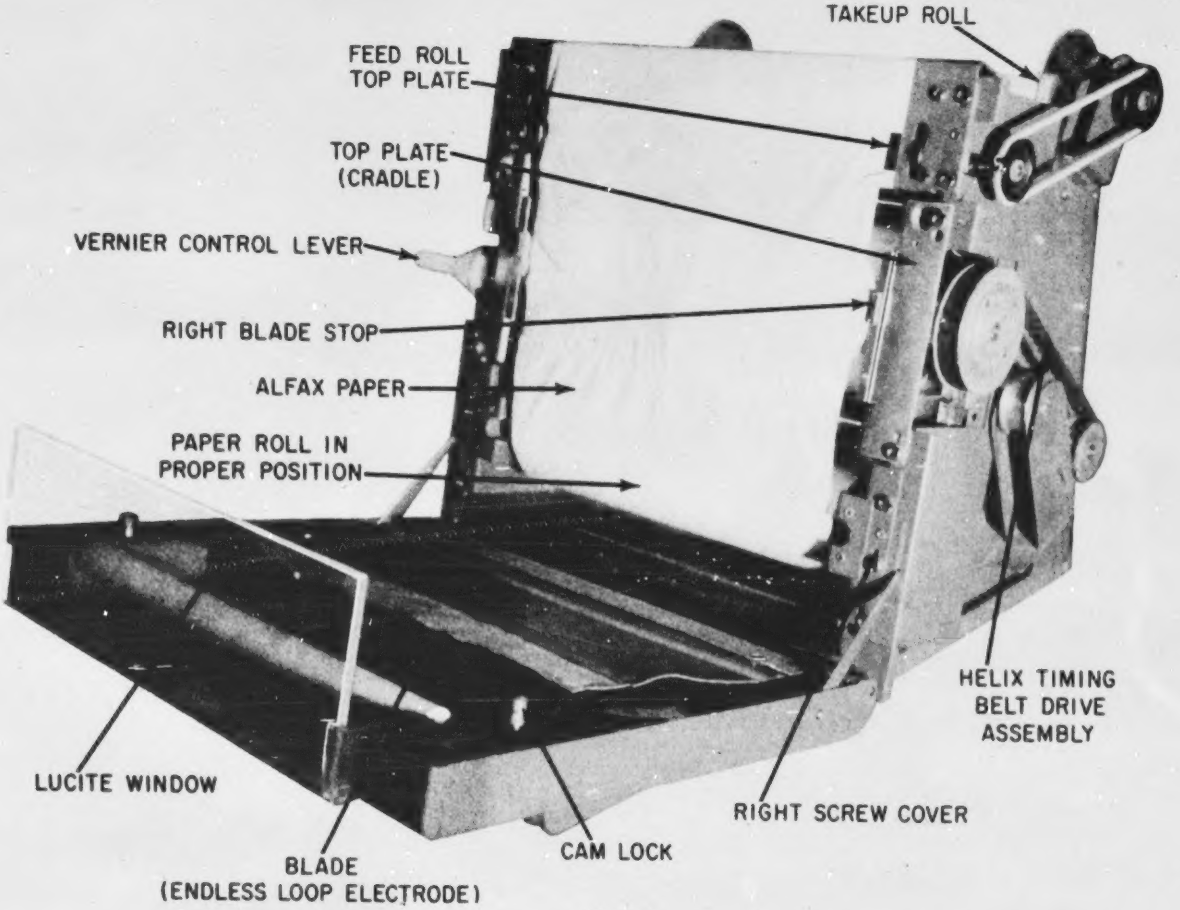
1. Clean the glass cover inside and out using clean cloth and alcohol.
2. Clean the reflector with a swab of cloth and alcohol. Using clean cloths each time, repeat the process until no dirt remains on the swabs.
3. Enter condition of reflector on checkoff list.
4. Stop and start to check for smoothness of operation.
5. Turn on drive motor and note motion of rotary mount. Check for noise.

Recorder

Taking care of the recorder (fig. 18-5) takes very little time. Like any piece of precision equipment, continued high level of operating results can be assured by following a few simple procedures.

Aerographer's Mates normally perform the following functions:

1. The helix and helix strip should be cleaned with each change of paper or after any extended idle period.
2. The paper supply tray should be cleaned before each new roll of paper is to be installed.
3. The drum and cradle assembly should be cleaned after every 1,500 to 2,000 hours of operation.
4. Install new rolls of paper in accordance with the basic manual of instructions. Make sure that the recorder cover is securely closed after installing the roll of paper so that it will not dry out. After installing the paper in the recorder, be sure that it is flat, smooth, and moist. This will assure the best marking condition.
5. If the degree marks are hazy, wider than usual, or fading out, open the cover of the recorder and make sure the blade is moving. If the blade is not moving, or is rough in certain areas, check the blade drive unit and replace the blade. In order to maintain clear and accurate marking, be sure that the blade is against its stops.


TAKEUP ROLL
FEED ROLL TOP PLATE
TOP PLATE (CRADLE)
VERNIER CONTROL LEVER
RIGHT BLADE STOP
ALFAX PAPER
PAPER ROLL IN PROPER POSITION
HELIX TIMING BELT DRIVE ASSEMBLY
LUCITE WINDOW
RIGHT SCREW COVER
CAM LOCK
BLADE (ENDLESS LOOP ELECTRODE)

AG.738

Figure 18-5.—Open view ceilometer recorder.

6. Replacing the helix and the blade (endless loop electrode) should be accomplished as needed in accordance with instructions in the technical manual for this piece of equipment.

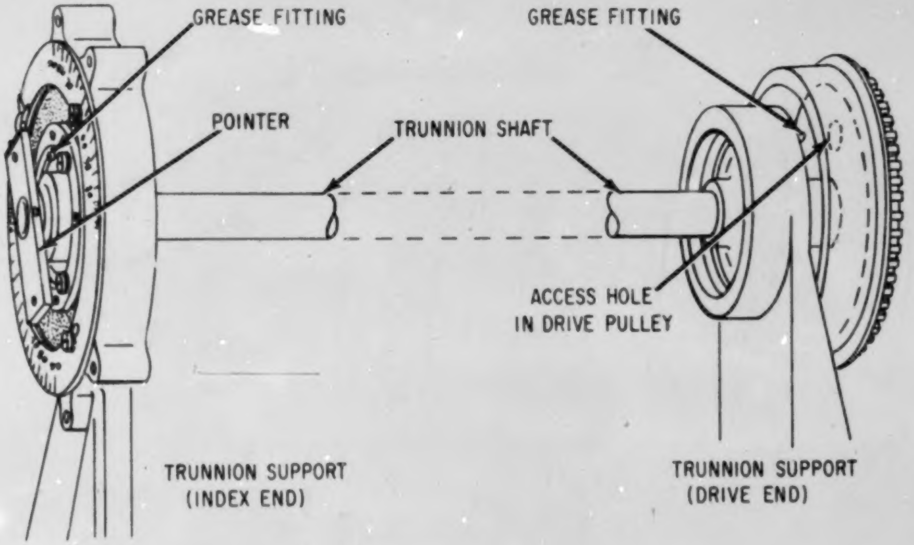
7. The only motor within the unit which will need lubrication is the paper feed drive motor. This motor should be lubricated (with the oil supplied) at the two oil holes at either end of the motor shaft. It should be lubricated every 6 months.

TROUBLESHOOTING AND REPAIR

Troubleshooting and repair other than that listed previously under the responsibilities for AG personnel are performed by ship, station, or contract electronic personnel. Details for accomplishing these functions are found in the Operation and Service Instructions Manual.

LUBRICATION

The trunnion shaft bearings and the drive pulley bearings of the projector require semi-annual lubrication. (See fig. 18-6.) No lubrication is required for the motors, or for any of the detectors components. Details for performing this lubrication are found in the Operation and Service Instructions Manual.


GREASE FITTING
GREASE FITTING
POINTER
TRUNNION SHAFT
ACCESS HOLE IN DRIVE PULLEY
TRUNNION SUPPORT (INDEX END)
TRUNNION SUPPORT (DRIVE END)

AG.739

Figure 18-6.—Projector lubrication diagram.

SEMIAUTOMATIC METEOROLOGICAL STATION AN/GMQ-14( )

The extensive use of the AN/GMQ-14 in measuring and recording surface environmental data makes its operating condition directly related to the accuracy of the data derived through its use. It is therefore essential that Aerographer's Mates be familiar with the use and care of this important meteorological measuring and recording system.

CALIBRATION

Calibration of the AN/GMQ-14( ) is performed by the Aerographer's Mate as described in the following paragraphs. Prior to commencing calibration, first wash and moisten the dewcel element as described in chapter 9, AG 3 & 2, NT 10363-D.

Dew Point Transmitter

A partial cut-away view of the dew point transmitter and dewcel assembly are illustrated in figure 18-7.

To calibrate the dewpoint transmitter, use the following procedure:

1. Prepare two containers of water. Insert an accurate thermometer in each one; keep one bath at 55°F and the other at 170°F, approximately.
2. Remove the dewcel from its protector, and remove the temperature bulb from the perforated guard.
3. Immerse the bulb in the lower temperature bath; stir well for about 1 minute. The temperature indicated by the thermometer may be assumed to be the temperature of the dewcel. Convert this reading to dewpoint temperature using the chart shown in figure 18-8.
4. Adjust the connecting link by means of the adjusting screw (fig. 18-7) so that the scale reading is the same as the dewcel reading.
5. Immerse the dewcel bulb in the higher temperature bath (170°F); stir for about 1 minute. Convert the thermometer to dewpoint reading and compare with reading now indicated on the data scale.
6. If scale reading is low, turn the calibration thumbscrew clockwise to adjust. If the scale reading is high, turn the calibration thumbscrew

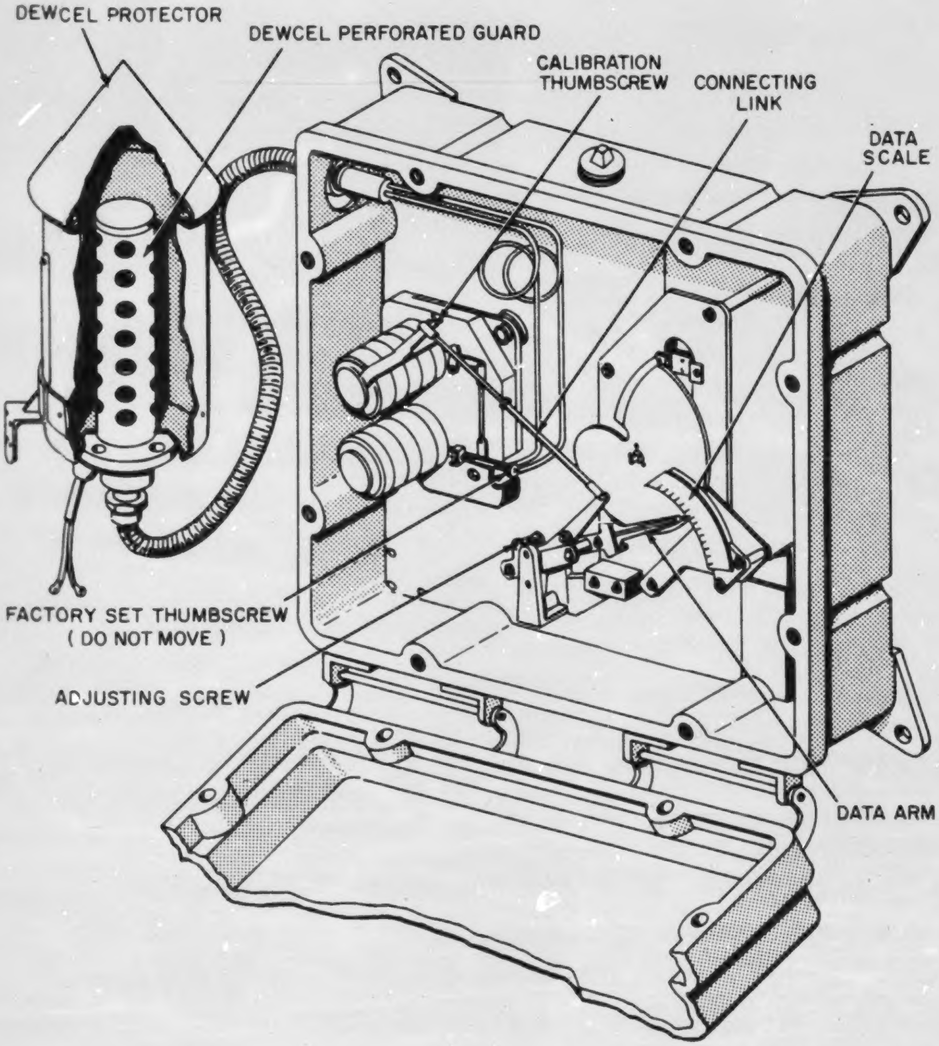

DEWCEL PROTECTOR
DEWCEL PERFORATED GUARD
CALIBRATION THUMBSCREW
CONNECTING LINK
DATA SCALE
FACTORY SET THUMBSCREW (DO NOT MOVE)
ADJUSTING SCREW
DATA ARM
AG.740

Figure 18-7.—Dewpoint transmitter and dewcel assembly.

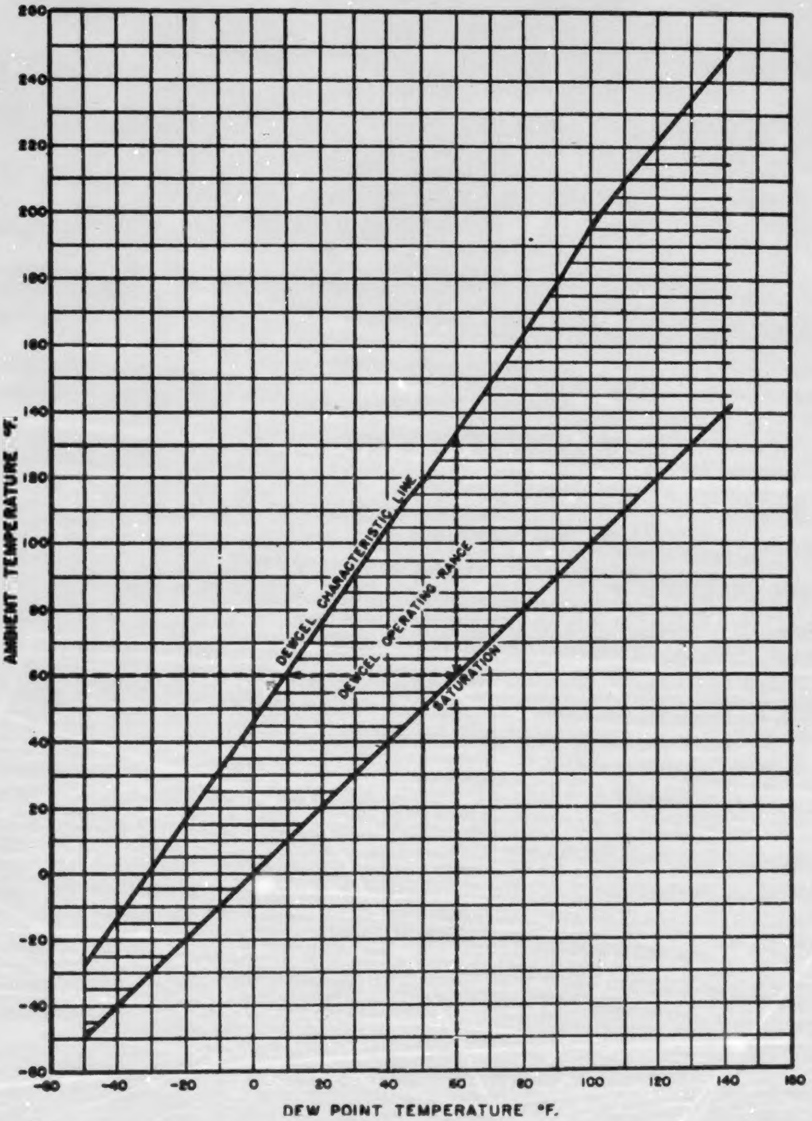
in the opposite direction. Adjust the reading of the data arm by means of the adjusting screw as mentioned in step 4.

CAUTION: Do not alter the position of the "factory set" thumbscrew; this is a compensation adjustment made at the factory and should not be moved.

7. Repeat step 6 using the first container (55°F) again. Continue this process using the two baths alternately until the error is eliminated from the scale reading.

Air Temperature Transmitter

The calibration of the air temperature transmitter is done in a similar manner as the calibration of the dewpoint transmitter, except that the temperature of the baths should be 10°F and 80°F, approximately, and there is no need for temperature conversion. The thermometer readings may be used directly.



AG.741

Figure 18-8.–Ambient temperature to dewpoint temperature conversion chart.

Dewpoint Receiver

To calibrate the dewpoint receiver one must first calibrate the dewpoint transmitter as described in the preceding paragraphs. Then proceed as follows:

1. Compare the readings on the transmitter indicator with the reading on the receiver indicator.

2. Adjust the receiver indicator reading to agree with the transmitter indicator reading.

Air Temperature Receiver

Calibration of the air temperature receiver is similar to that of the dewpoint receiver, except that the air temperature receiver indicator is compared with the air temperature transmitter indicator.

MAINTENANCE

The maintenance that Aerographer's Mates are required to perform on Semiautomatic Meteorological Station AN/GMQ-14( ) is discussed in chapter 9 of Aerographer's Mate 3 & 2, NavTra 10363-D.

The tests that are performed on the electronic components of this equipment are made by qualified technical personnel only.

WEATHER TELEVISION SYSTEMS

The consoles of the Weather Television System AN/GMQ-19( ) currently in use and its replacement, Weather Television System AN/GMQ-27( ) are illustrated in figures 18-9(A) and (B).

There are a number of modifications which have been incorporated into the newer AN/GMQ-27. However, these changes are primarily improvements in solid state circuitry, camera design, etc. Maintenance procedures performed by contract or station electronic personnel have changed to some degree. The minor preventive maintenance required of the Aerographer's Mate remains relatively unchanged and will be discussed later in this section. The system arrangement and related functions have remained relatively stable and are briefly presented in the following paragraphs.

SYSTEM ARRANGEMENT

All components of the central weather station are mounted within or on the metal cabinet assembly. Two rear access doors are provided which are hinged and will lift off. The light table on the right side of the console contains six fluorescent lamps beneath a glass plate and diffusion glass cover. Two 150 watt floodlights are provided for opaque illumination when not using the light table for transparent illumination.

The camera assembly with its component parts is mounted rigidly to the top of the cabinet over the light table. A covered housing protects the lens zoom and focus motors.

The recorder is located just above the video screen in the left section of the console. It is a single channel, 24 hour, magnetic tape recording and reproducing unit designed to record from either a microphone, a telephone, or a direct line. It will record continuously for 24 hours (plus a 15 minute overtime allowance) without tape change. Cranks are provided on the reel spindles for manually advancing or rewinding tapes. The tape motor is driven in the forward direction only and must be rewound manually. A portable tape demagnetizer is provided with the unit for degaussing of the magnetic recording tape. The unit will degausse a 2-inch roll of magnetic recording tape in 5 to 10 seconds.

The audio system permits two way voice communications between the central weather station and the remote stations. An audio and video selectivity system is provided to permit briefing any remote station in private, or the entire system simultaneously at the briefer's option.

Control panels on the console contain all the necessary switches and controls for operation of the system.

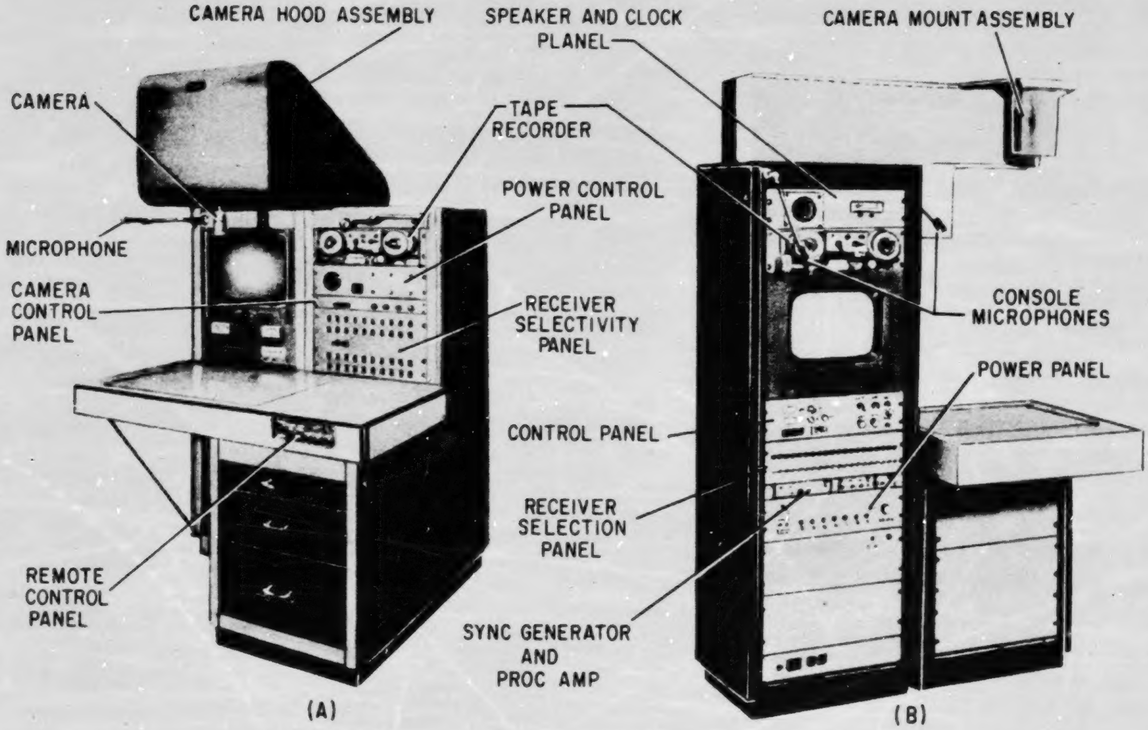
The television monitors used at remote locations consist of a 21-inch television monitor mounted in a metal case assembly. Normal television controls pertaining to contrast, brightness, etc., are provided on the front panel.

CHECKING OPERATION

The weather television system should be checked daily for proper operation. Both the video and audio portions should be activated, with checks of reception at remote stations being made. If faulty operation is noted during the daily check, proper maintenance personnel should be notified.

MAINTENANCE

Contract employees or station electronics personnel perform the required electrical and electronic maintenance on weathervision systems. However, there are a number of minor


CAMERA HOOD ASSEMBLY
SPEAKER AND CLOCK PLANEL
CAMERA MOUNT ASSEMBLY
CAMERA
TAPE RECORDER
POWER CONTROL PANEL
MICROPHONE
CAMERA CONTROL PANEL
RECEIVER SELECTIVITY PANEL
CONSOLE MICROPHONES
POWER PANEL
CONTROL PANEL
RECEIVER SELECTION PANEL
REMOTE CONTROL PANEL
SYNC GENERATOR AND PROC AMP
(A)
(B)

AG.742

Figure 18-9.—(A) Weather Television System AN/GMQ-19( )console; (B) Weather Television System AN/GMQ-27( ) console.

preventive maintenance tasks and inspections which may be performed by the Aerographer's Mate. These are limited to the inspections, checking of switches, and cleaning described in the following paragraphs.

Chassis Assembly Components

The components of all chassis assemblies should be inspected for dirt, rust, or corrosion on a monthly basis under normal conditions. If the equipment is located under exceptionally hot, humid, or windy and dusty conditions, or in the proximity of salt water, these inspections should be conducted weekly.

If cleaning is required, remove grease and oil from the chassis and cabinets using cleaning solvent, Federal Specification P-S-661, or equal. To clean painted panels use cleaning compound Military Specification MIL-C-18687, and warm, clean water. Rinse with clean water.

CAUTION: If solvent is used be sure the area is well ventilated. Do not inhale solvent vapors or allow solvent to come in contact with the skin. Keep solvent away from open flame.

Television Viewers

The glass on the front of the television viewers should be cleaned weekly under normal conditions; if other than normal conditions exist, they should be cleaned daily.

If cleaning is required use a soft cloth soaked in aliphatic naphtha Federal Specification TT-N-95A. Polish with a clean dry cloth.

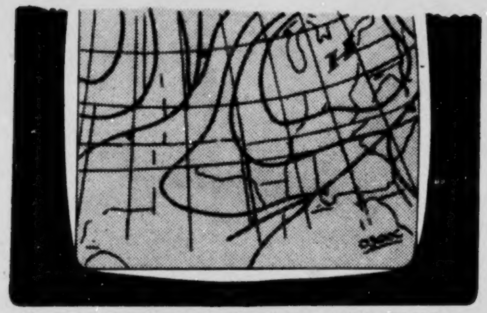
Console Light Table

The plastic cover on the console light table should be cleaned weekly unless required more often due to windy and dusty conditions. Cleaning should be accomplished by washing with a soft cloth and warm, clean water and then polishing with a soft dry cloth.

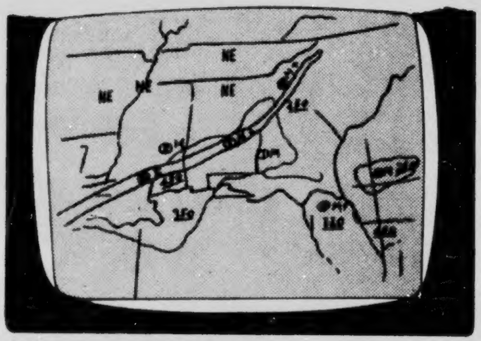
The lamps in the console light table should be checked daily to insure that they are operating properly.

WEATHERVISION PRESENTATION

Weathervision systems have proven to be a valuable tool in the weather briefing situation at stations where they have been installed. Figure 18-10(A) and (B) shows some of the types of weather data that can be presented on weathervision equipment.



(A) SURFACE WEATHER MAP WITHOUT STATION MODELS



(B) RADAR SUMMARY CHART

AG.743

Figure 18-10.—Typical weather vision transmissions. (A) Surface weather map; (B) radar summary chart.

METEOROLOGICAL RADAR

Radar and weather are very closely allied. The use of radar has provided meteorologists with a tool which permits the collection of atmospheric data under conditions when the more orthodox methods fail. On the other hand, a knowledge of atmospheric conditions provides the radar operator with information vital to the accuracy of his results.

The word radar was derived from the phrase "radio detection and ranging." Fundamentally, all radar depends upon the emission of a short, sharp pulse of electromagnetic energy in a given direction. This pulse on intercepting a target is scattered in all directions. That part of the energy which is scattered in the direction of the radar is picked up by the radar antenna, producing an echo or "blip" on the receiver scope. The time taken for the pulse to cover the path to the target and return is a measure of the range to the target. Azimuth and elevation of the target are determined from the direction in which the pulse is emitted and returned.

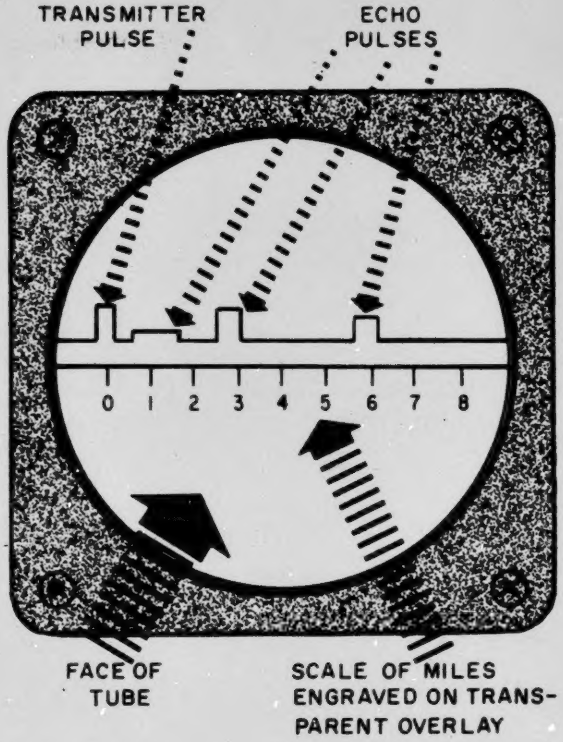
Since the electromagnetic pulse travels with the speed of light, range is determined by a timing procedure. At the instant the pulse leaves the antenna, a timing device starts counting. When the return echo reaches the antenna, the counting is stopped. The distance to the target is equal to one-half the distance traveled by the pulse.

RADAR INDICATORS

The purpose of the indicator (cathode ray tubes) in a radar is to display information to the Aerographer's Mates about the range, bearing, and elevation of surrounding targets. In general, information is presented piecemeal, and a single viewing of an indicator gives only a small part of the picture. Different types of radar scans are employed to display wanted information from returning echoes on the different indicators.

A-Scan Presentation

The simplest type of display is the A-scan presentation, which is illustrated in figure 18-11. The display is one which gives return signal intensity against range. This display shows only the range to a target.


TRANSMITTER PULSE
ECHO PULSES
0
1
2
3
4
5
6
7
8
FACE OF TUBE
SCALE OF MILES ENGRAVED ON TRANSPARENT OVERLAY

AG.744

Figure 18-11.—General appearance of A-type scan presentation as it would appear on the range indicator CRT.

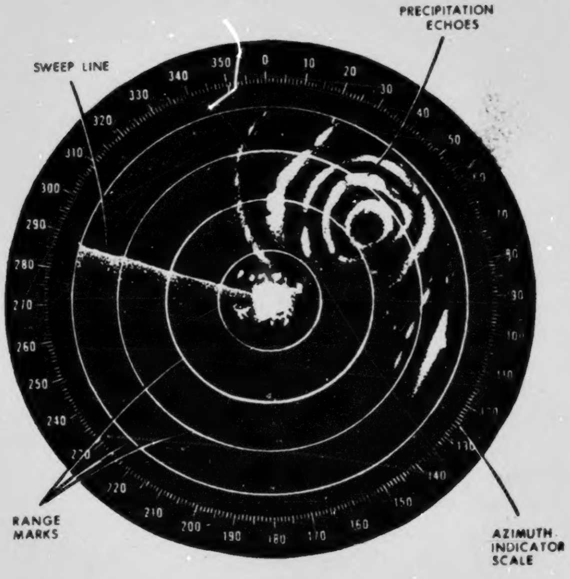
If the target is a single, discrete object such as an aircraft, the bearing of the target can be found by moving the antenna horizontally to the position of maximum signal intensity (highest pip). Changing the angle of elevation of the antenna to get maximum signal intensity provides information concerning the elevation of the target. The horizontal and vertical limits of a rainstorm can be found in much the same way by noting the positions of the antenna at which the return drops below a detectable level.

R-Scan Presentation

The R-type scan is almost identical to the A-type. The signal intensity is given as a function of range. Essentially, the difference between the two is that the A-scan presentation displays the total range starting at 0 range on the left of the scope and ending at any one of several ranges selected; the R-scan presentation isolates a portion of the range scale and expands it across the entire face of the tube. Thus, perhaps only the range between 20 and 40 miles is displayed, or any other range interval may be selected.

PPI-Scope

Another type of display commonly used in radar is termed plan position indication (PPI). This display makes it possible to read range and bearing information simultaneously. A diagram of a basic PPI-scope is shown in figure 18-12.


PRECIPITATION ECHOES
SWEEP LINE
RANGE MARKS
AZIMUTH INDICATOR SCALE
0
10
20
30
40
50
140
150
160
170
180
190
200
210
220
240
250
260
270
280
290
300
310
320
330
340
350

AG.745

Figure 18-12.—Diagram of PPI-scope.

In essence, with the PPI-scope, the sweep starts at the center of the tube instead of at one edge as it does on the A-scope. As the antenna rotates around a 360° circle, the sweep rotates simultaneously. The intensity of the sweep display is modified by the presence of a return signal. Thus, the position of a target is indicated at the correct azimuth and range by a bright spot on the display tube. A maplike picture is thus produced on the tube as the antenna rotates in the azimuth plane.

RHI-Scope

The range height indicator (RHI) is another scope occasionally found on radar equipment. Its function is not unlike that of the PPI except that when this scope is used, the azimuth of the antenna is kept fixed while the antenna moves from viewing at 0° elevation to a predetermined elevation. The sweep always starts at the lower left-hand side of the cathode tube and moves with the antenna as it oscillates vertically.

Range Markers

Each of the conventional scopes provides for some direct indication of range by display of range markers. The range markers may be turned on or off and may be varied. When the scope is set for 10- or 25-mile display, range markers at 1- or 5-mile intervals are convenient. At 100, 200, or 400 miles, 25- or 100-mile markers are frequently used.

RADAR EQUIPMENT

Meteorological radar provides a unique means of obtaining meteorological data for use by the forecaster when issuing warnings and other environmental forecasts. As weather requirements continue to expand and change, new designs and mofifications to meteorological radar will continue to appear in the effort to improve and keep pace.

The Meteorological Radar Set AN/FPS-106 is the newest model of radar designed for meteorological use. A limited number of these have been manufactured and put into operation at this time. The AN/FPS-81 meteorological radar is the most common set in use in the Navy today. There are however, older models, FPS-41, and FPS-68, which are still in operation and may be encountered. Since there are many similarities between the FPS-41, FPS-68, and FPS-81, and the FPS-106 is in such limited use, only the FPS-81 will be discussed at length in this manual; brief mention will be made of the older models where notable differences exist.

Purpose of Equipment (AN/FPS-81)

Meteorological Radar Set AN/FPS-81 is a ground-based radar system used to establish the geographic locations of storm centers relative to a fixed base reference site. The equipment is capable of detecting storm centers within a radius of 200 nautical miles. Radar echo signals from concentrations of high moisture content are presented on cathode-ray tube (CRT) indicators in such a manner as to convey the positions of storms in terms of azimuth, slant range, and elevation (height). The radar echo signals may be displayed with iso-echo contouring, if desired, to provide maximum clarification of storm configurations.

Limitations

The antenna may be made to scan manually in both planes (azimuth and elevation) simultaneously, or it will scan automatically in one plane while manual search is conducted in the other. The antenna will not scan automatically in both planes simultaneously. Radar targets will not normally be visible at less than a 1-mile range.

Component Parts

Meteorological Radar Set AN/FPS-81 consists of an antenna assembly, a receiver-transmitter-modulator (RTM) assembly, an indicator console, and a remote indicator assembly. (See fig. 18-13.) The antenna may be separated from the RTM assembly by a maximum distance of 100 feet; the indicator console may be located up to 2,600 feet from the RTM assembly; and the remote indicator assembly may be operated 5,300 feet from the indicator console. The major portion of the system circuitry is contained in modular units of the plug-in type to simplify and expedite maintenance procedures.

ANTENNA ASSEMBLY. The antenna group (fig. 18-14(D)) consists of a spun aluminum parabolical dish, measuring 8 feet in diameter; a horn-shaped reflector feed; and an azimuth pedestal. Radar illumination of target areas up to 200 nautical miles away is provided by the antenna which directs and concentrates the high-level RF (radiofrequency) energy into a

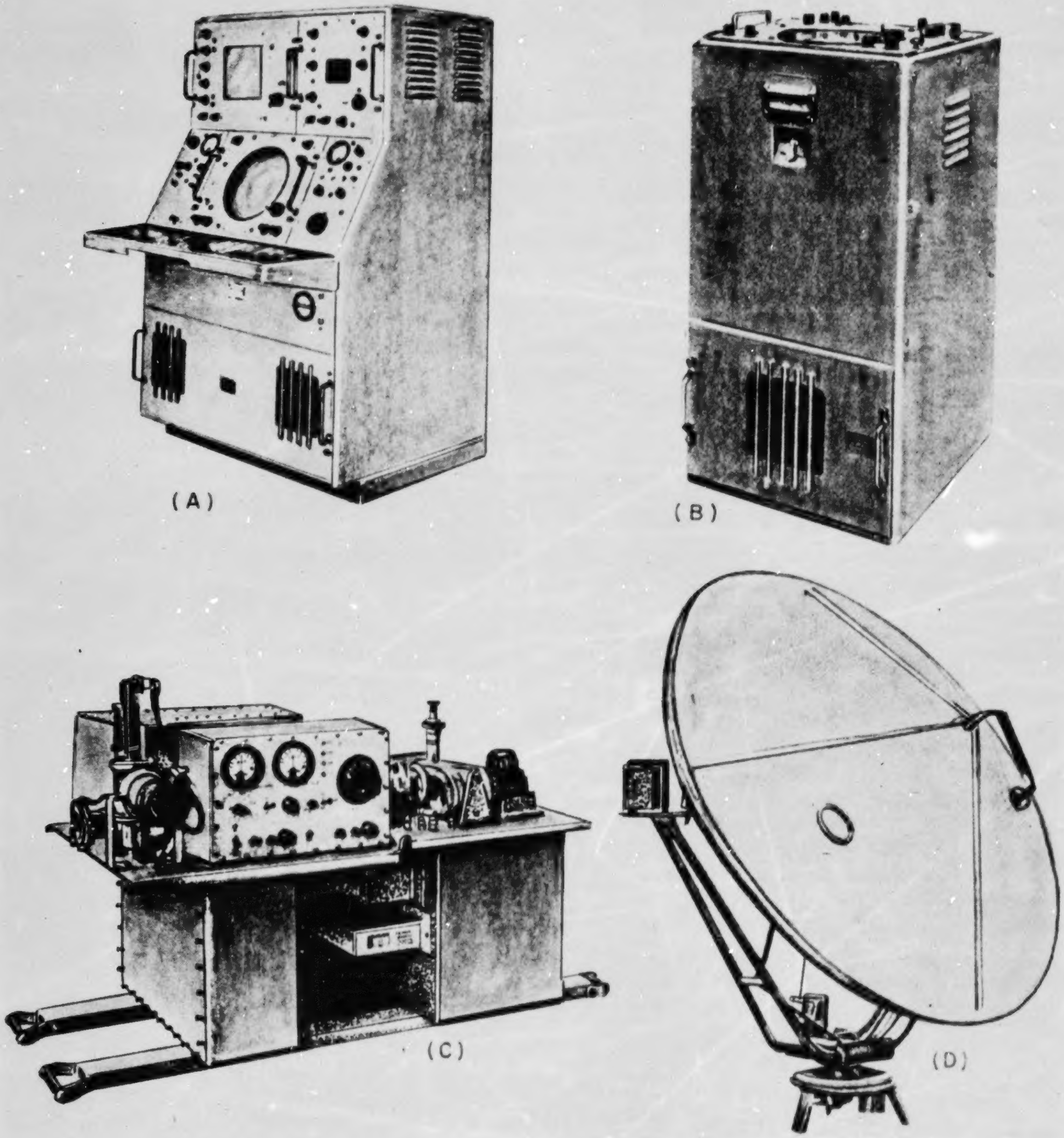


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Figure 18-13.–Meteorological Radar Set AN/FPS-81. (A) Indicator Group OA-3871/FPS-81; (B) Indicator Group OA-3872/FPS-81; (C) Receiver-Transmitter, Radar RT-658/FPS-81; (D) Antenna Group OA-3870/FPS-81.

narrow beam. The radiated beam is a burst of energy generated in the frequency band of 5,450 to 5,650 megahertz and pulsed 324 times per second for a duration of 2 microseconds. During the unpulsed portion of the pulse repetition interval, radar echoes are returned from target areas to be detected by the antenna and fed to amplifying, timing, and indicating circuits throughout the system.

Control of the antenna may be accomplished manually or automatically. Each mode of antenna scanning permits separate control of azimuth motion. In automatic azimuth scanning, the antenna scans 360° at a constant rate of 5 rpm. In automatic elevation scanning, the antenna nods continuously from +60° to −2°.

RECEIVER-TRANSMITTER-MODULATOR (RTM) ASSEMBLY.–The receiver-transmitter assembly (fig. 18-14(C)) contains the electronic components required to transmit and receive the RF signals. Controls are provided for the adjustment of output power and transmitter frequency. The magnetron oscillator in the transmitter-modulator section of the RTM assembly generates RF energy bursts which are radiated by the antenna system.

INDICATOR CONSOLE.–The indicator console, from which the entire system may be controlled and monitored, houses the range/height indicator (RHI), the plan position indicator (PPI), and the Range Indicator as illustrated in figures 18-13(A) and 18-14.

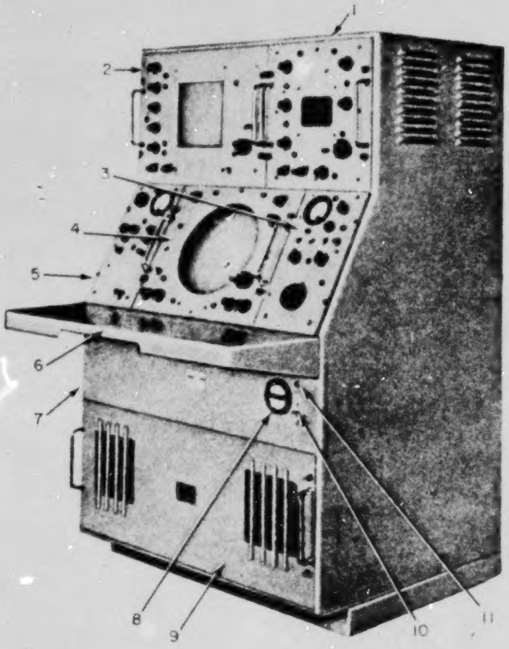
A power panel on the rear of the indicator console, mounted between the PPI and range indicator assemblies, accepts 115-volt, 60-cycle, a-c power for distribution to the entire system.

PLAN POSITION INDICATOR (PPI).–The PPI, located in the center portion of the indicator console, presents range and azimuth information in a plan view (as seen from zenith directly over the radar site).

RANGE/HEIGHT INDICATOR (RHI).–The RHI indicator occupies the left side of the indicator console and presents range and height information. Range information is displayed along the horizontal axis of the RHI display, while height data appears along the vertical axis. Selection may be made of two height scales, 0-40,000 feet, or 0-80,000 feet.

The RHI assembly contains the circuits which resolve the height component of a selected target from inputs of slant range and antenna elevation angle. The height video signal displayed on the RHI is automatically corrected for earth curvature at all ranges within the capabilities of the radar set. Antenna scanning in elevation is controlled from the RHI.

RANGE INDICATOR.–The range indicator, housed in the upper right-hand section of the indicator console, features a standard A-scan 5-inch cathode-ray tube, which also contains a



AG.747

1. **Range Indicator IP-644/FPS-81**
2. **Range Height Indicator IP-643/FPS-81**
3. **Right-hand control panel**
4. **Azimuth Range Indicator IP-642/FPS-81**
5. **Left-hand control panel**
6. **Desk**
7. **Cabinet**
8. **Elapsed-time Meter M7201**
9. **Amplifier-Power Supply AM-3306/FPS-81**
10. **Power Circuit Breaker CB7201**
11. **Phone jack**

Figure 18-14.—Indicator console, AN/FPS-81, assembly locations.

ranging function. Range information (A-scan) is displayed along the horizontal axis of the CRT display. The R-presentation along the horizontal axis consists of a continuously variable range strobe which may be superimposed manually over the target echo pulse to provide precise ranging data. The range strobe, generated within the range indicator assembly, represents a range interval 5 miles in depth. By actuating a switch to select the R-scan, this range interval may be expanded to fill the range indicator screen, permitting a more detailed analysis or photography of the echo trace selected by the range strobe setting. The selected range strobe also drives a decimal digital device on the front panel which presents the actual range in nautical miles.

Sweep timing circuits permit selection of sweep ranges of 30, 60, 120, and 200 nautical miles. Range markers may be displayed along the horizontal display baseline at intervals representing 5-, 10-, 25-, and 50-mile range increments. The range indicator includes self-checking circuitry which enables the operator to check the accuracy of the range marker signals for all ranges simply and quickly.

REMOTE INDICATOR. The remote indicator assembly, which may be located up to 5,300 feet from the indicator console, provides a repeat display of the information presented by the local indicator at the indicator console Video and synchronizing signals for the remote indicator are supplied by the console local indicator assembly. The sweep trace on the remote indicator rotates in synchronism with the antenna scanning motion in azimuth.

PREVENTIVE MAINTENANCE

Radar equipment is complex electronic equipment; therefore, servicing and maintaining of this equipment is the responsibility of the electronics technicians. Aerographer's Mates' responsibilities only include keeping the exterior of the equipment clean and promptly reporting internal difficulties to the responsible technicians.

METEOROLOGICAL RADAR SET AN/FPS-41

The components are essentially the same as those for the AN/FPS-81, except that the transmit-receive units are in the same assembly with the reflector.

Other important differences are that the remote indicator can only be mounted a maximum distance of 500 feet from the main console; the range is from 250 nautical miles to 1/2 mile; the indicator console is divided into six panels instead of three on the AN/FPS-81, and the controls are divided among the upper and lower panels, whereas on the AN/FPS-81 the controls are located on the panel with the three scopes.

METEOROLOGICAL RADAR SET AN/FPS-68

The AN/FPS-68 meets basically the same contract specifications as the AN/FPS-81, but is not as compact and modern in design.

SAFETY PRECAUTIONS

High voltage is used in the operation of all radar. Extremely dangerous voltages exist in the receiver, transmitter, modulators, main console, and remote indicator. Death on contact may result if operating personnel fail to observe safety precautions. Only authorized operation of equipment should be carried out by the operator. All operators must know how to secure main power, both remote and locally, to the set in use. All operators must know how to apply artificial respiration and how to contact immediate medical aid.

FACTORS AFFECTING RADAR PERFORMANCE

There are many factors, or elements, that affect efficient radar performance, not all of which are completely understood. Many of the limitations of the equipment are a result of radar design. If you recognize these limitations and their causes, you will be better able to determine the type of performance to be expected from your radar.

One important factor is the radar operator's knowledge of this equipment. He must know the maximum and minimum ranges at which he can expect to pick up the weather echoes and the range and bearing accuracy of the gear. These characteristics can be determined for your particular set from the applicable technical manual. In this section of the chapter only those factors which generally affect all radar equipment are discussed.

Frequency

Frequency/wavelength determines the efficiency with which radar energy is scattered by a target. The closer the wavelength of the radar energy is to the size of the target, the more efficiently the target scatters the radiation. Targets not only scatter radar energy, they also absorb it. This is why some wavelengths do not penetrate some targets. Loss of radar energy as a result of two factors, scattering and absorption, is known as attenuation. For meteorological phenomena, the targets are the water droplets or ice particles in the clouds and areas of precipitation. A 20-cm radar would detect the large raindrops of a thunderstorm, but would not detect the small raindrops and cloud droplets. On the other hand, a 3-cm radar would detect the small droplets, but the energy would be so rapidly dissipated by absorption and scattered so effectively by the small droplets on the near side of the cloud that no energy would penetrate to the back side.

Pulse Length

The pulse length is a factor in determining the maximum and minimum range at which a target can be detected and the power of resolution (degree to which details can be distinguished).

1. Minimum range. If a target is so close to the radar site that the reflected energy returns to the antenna before the trailing edge of the pulse has left the transmitter, it will not be detected because it will return while the receiver is not listening.
2. Maximum range. The reflected radar signal must be above the minimum energy level to which the radar will respond. Long pulses contain more energy than short pulses; therefore, distant targets and weak targets are more likely to be detected.
3. Resolution. Whether two targets located on the same azimuth from the radar site will be detected as one echo or as two depends on the pulse length. All the reflected energy from the closest echo must have returned to the antenna before the reflected energy from the second echo reaches the antenna, if the two are to appear as separate echoes.

Pulse Repetition Frequency (PRF)

The PRF determines the maximum measurable range (MMR) of the radar. Ample time must be allowed between pulses for an echo to return from any target located within the maximum workable range of the system. Otherwise, returning echoes from the more distant targets will be blocked by succeeding transmitted pulses. This necessary time interval fixes the highest PRF that can be used. The lower the PRF, the greater the range. However, the PRF must be high enough so that sufficient pulses hit the targe and enough are returned.

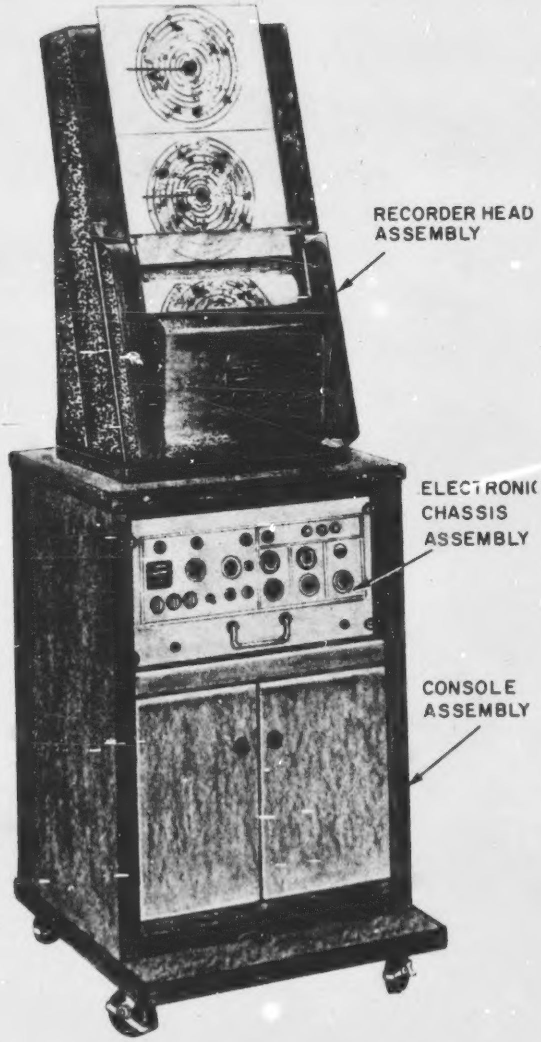
Beam Width

Beam width is determined by the size and shape of the antenna. With a wide beam there is more chance of detecting a target, but there is less chance of locating it accurately in azimuth and elevation. The more concentrated the beam, the greater the range capabilities for a given amount of transmitted power.

RADAR FACSIMILE RECORDER AN/GMH-6( )

The Radar Facsimile Recorder AN/GMH-6( ) is illustrated in figure 18-15.

This recorder is a remote picture printing device used to record weather data transmitted from the radar/transmitter site where the transmitting device horizontally scans a plan position indicator (PPI) radar scope. The data is then transmitted via telephone line to the recorder. Ther recorder provides a hard copy printout of the weather pictures, including "data insert information" on a continuous roll of electrolytic paper. The data insert information consists of automatic printing of a time-date code, indicating the day of the year and the time of the day that the particular picture was received and printed.



AG.748

Figure 18-15.—Radar Facsimile Recorder AN/GMH-6( ).

The recorder has controls that determine the frequency at which consecutive groups of pictures are printed. A group may consist of 1, 2, or 3 pictures. The period between the groups may be 5, 10, 20, 30, 60, or an infinite number of minutes. A period in this case refers to the length of time from the end of the last picture of a consecutive group to the beginning of the first picture in the following consecutive group.

Controls are also provided for the adjustment of printing quality, such as degree of contrast and whiteness. Other controls set the time and data information printed on each weather picture.

As illustrated in figure 18-15, the facsimile recorder consists of the electronic chassis, and the recorder head (printing display area and paper takeup) which are mounted in and on, respectively, one metal console. The console rests on four swivel casters, to permit mobility. The two front casters can be locked to fix the console in a stationary position. The electronic chassis is a slide-mounted pullout drawer which provides for easy removal, servicing, cleaning, and testing of the chassis. The electronic chassis contains all operating controls, fuses, and indicators for the facsimile recorder, except the paper take-up switch and fuse, the paper supply indicating light, and the paper fast feed switch, which are located on the Recorder Head.

OPERATION

Prior to operation of the Radar Facsimile Recorder, personnel should refer to the publication NAVAIR 50-30GMH6-1. Complete operating instructions are contained in this publication.

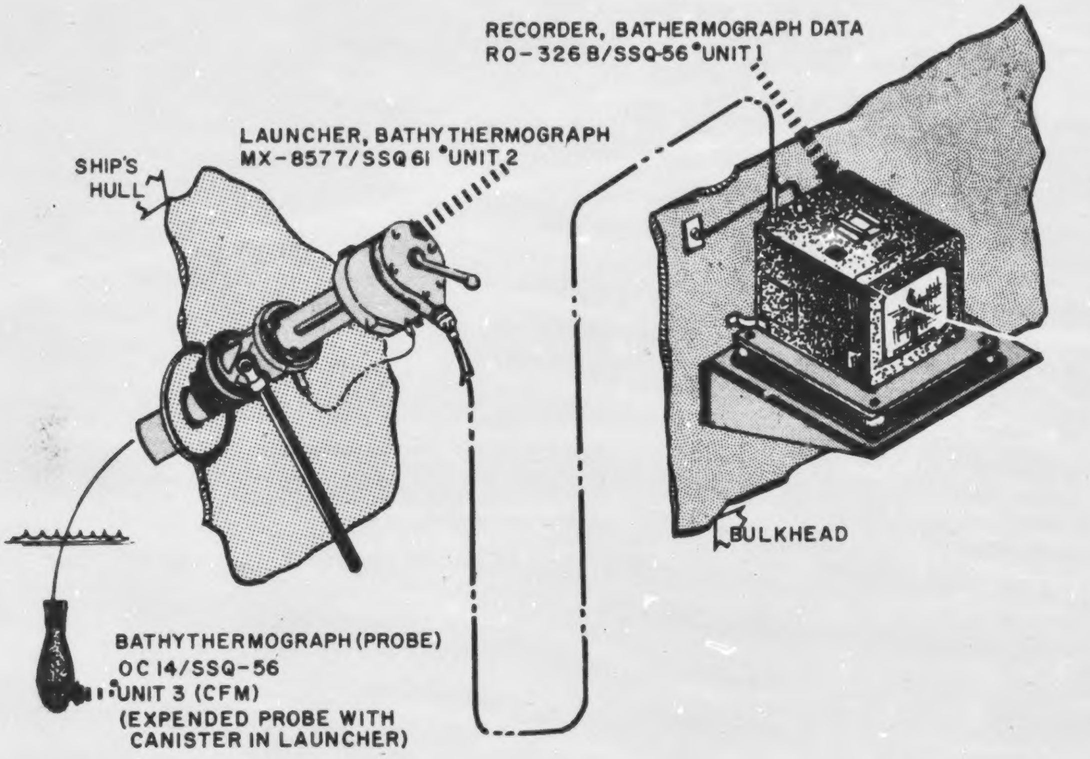
MAINTENANCE

Servicing and maintaining of this equipment are the responsibilities of trained electronics personnel As with other complex electronic equipment the maintenance duties of the Aerographer's Mate are limited to keeping the exterior of the equipment clean and promptly reporting any internal difficulties to the responsible parties.

OCEANOGRAPHIC INSTRUMENTS

BATHYTHERMOGRAPH SYSTEMS

The operation of bathythermograph systems such as the AN/SSQ-61 illustrated in figure 18-16 was described briefly in AG 3 & 2, NT 10363-D. Additional details and modifications are discussed in the following paragraphs.


RECORDER, BATHERMOGRAPH DATA
RO-326B/SSQ-56 *UNIT 1
LAUNCHER, BATHYTHERMOGRAPH
MX-8577/SSQ61 *UNIT 2
SHIP'S HULL
BULKHEAD
BATHYTHERMOGRAPH (PROBE)
OC 14/SSQ-56
*UNIT 3 (CFM)
(EXPENDED PROBE WITH
CANISTER IN LAUNCHER)

AG.749

Figure 18-16.—Bathythermograph Set AN/SSQ-61, relationship of units.

This bathythermograph (BT) system will automatically obtain and record ocean temperature data in virtually any sea state without altering the ship's normal operation. It will obtain a complete temperature profile of the sea to a depth of 1,500 feet in 90 seconds. The AN/SSQ-61 is the most current in the series of BT systems. Its predecessors were the AN/SSQ-56, AN/SSQ-56A, and the AN/SSQ-60. Each system incorporates electrical and mechanical improvements over the older model. Differences between the various sets lie only in the recorder and the launcher; all sets use the same XBT probe (OC-14/SSQ-56). Table 18-2 lists the set components and major differences along with the applicable technical manuals.

Any combination of recorder and launcher may be utilized with the XBT. Minor differences affecting installation, operation, and maintenance may be readily determined by reference to the applicable manuals.

Operating Malfunctions

As with all equipment, the ability to detect and correct potential problems by the operator may save major expense and equipment outage. Table 18-3 lists some of the potential problems and correction procedures on the AN/SSQ-61 which may be performed by the operator.

Maintenance

The maintenance which may be performed on the BT system AN/SSQ-61 by the operator is limited to that presented in the following paragraphs. Although many similarities exist between systems as stated previously, the applicable technical

Table 18-2.–Equipment Similarities

Set	Recorder	Differences from previous recorder	Launcher	Differences from previous launcher	Technical manual
AN/SSQ-56	RO-326/ SSQ-56	Earliest model. Chart paper moves from top to bottom. Bridge, trigger logic and power supply on two circuit boards, hardwired to harness.	MX-7594/ SSQ-56	Earliest model, deck mounted.	NAVSHIPS 0967-225-6010
AN/SSQ-56A	RO-326A/ SSQ-56	Major mechanical and electrical changes. Recorder instrument assembly inverted for improved chart display. Bridge, trigger logic and power supply combined on single connector-mounted circuit board. Test panel and switches added.	MX-7594A/ SSQ-56	Minor mechanical changes in stanchion, breech adapter and launch tube.	NAVSHIPS 0967-305-6010
AN/SSQ-60			MX-8416/ SSQ-60	Major mechanical changes. Launcher installed through hull rather than on deck. Uses heavy muzzle protector.	
AN/SSQ-61	RO-326B/ SSQ-56	Minor mechanical and electrical changes. Added power switch and servo muting.	MX-8577/ SSQ-61	Major mechanical changes in launch tube and mounting. Stanchion and muzzle protector eliminated. Launch tube changed to urethane material.	NAVSHIPS 0967-333-6010

Table 18-3.–Malfunction Chart
(Operator Level)

This table describes the simplest malfunctions, their symptoms, causes and corrective actions. These are generally considered on-site procedures and are within the capability of the operator.

SYMPTOM	CAUSE	CORRECTION
During measurement run stylus makes erratic excursions to the right end of the chart paper (high temperature end)	Leak from probe wire to the salt water.	Launch new XBT.
During measurement run stylus makes excursion all the way to left (low temperature end) of chart paper and remains there.	Complete wire break.	Launch new XBT.
During measurement run stylus makes erratic excursions to the left (low temperature end) of the chart paper.	Contamination between pin B of breech and canister.	Breech pins should be wiped thoroughly and new XBT inserted for another launching.
	Excessive salt water in breech or receiver assembly causing conductive path across contact end of canister.	Wipe breech and receiver thoroughly and insert new XBT for another launching.
Recorder completely inactive, indicators not illuminated.	POWER switch off or fuse blown.	Replace blown fuse. Locations are on top rear of Recorder interior. If fuse blows again notify the maintenance technician.
Calibration line marked during Check/Run Mode full outside 61.8° - 62.2° band.	Recorder out of Calibration.	Calibrate following the procedure described in the equipment technical manual.

manual should be checked prior to commencement of maintenance procedures.

LAUNCHER. The operator at the launcher must insure that no scran wire from the launching remains in the launch tube when the ball valve is closed, and that the breech and tube are kept clean and free of contaminants. Contaminants may be removed with fresh water and cloth. The launch tube must be inspected weekly for nicks or burrs which might damage the fine XBT wire during launching. The tube must be smooth over its entire interior surface and around the opening.

CALIBRATION CHECK.–Before a series of XBT launchings, or at weekly intervals when XBT's are launched daily, a system calibration check should be made by the operator. The calibration check is performed with the use of the test canister provided with the system and following the procedure as outlined in the technical manual.

BT RECORDER. Remove, install, and aline chart paper as described in the applicable technical manual. The procedure to be followed involves a number of steps which must be followed precisely and since the technical manual should accompany the equipment, it is considered unnecessarily repetitious to include these steps here.

BATHYTHERMOGRAPH PROBE.–No preventive maintenance is required for the probe and canister.

PORTABLE RADIATION THERMOMETER PRT-4( )

The PRT-4( ) system is a noncontact passive instrument that measures the radiation from remote water surfaces, clouds, backgrounds, and other targets filling the field of view of its optical system. It is particularly suited for airborne use, where its fast response and high resolution power permit the mapping of thermal contours and surface currents. The operation of the PRT-4( ) is presented in AG 3 & 2, NT 10363-D.

MAINTENANCE

Other than occasional visual checks and cleaning of the lens surface, no routine maintenance of the system is necessary. The visual checks should be conducted annually, or at intervals as required by the service environment and extent of use. During these checks, particular attention should be given to cables and connectors, front panel components (switches and indicators), and mounting hardware.

Cleaning the Optical Lens

The front window may require cleaning from time to time. (This is most often indicated by deterioration in system performance.) However, excessive cleaning is undesirable because the window may become permanently scratched or marred. Therefore, before undertaking to clean the window, check the system performance under a known set of operating conditions. If deterioration is evident, inspect the window surface for evidence of a haze of dust; if present, remove with a clean, optical-grade camel's hair brush, using a flicking motion. Do not scrub, and do not use lens tissues or cleaners. If further cleaning is required, the instrument should be returned to an authorized repair facility.

LUBRICATION

All moving parts in the Portable Radiation Thermometer PRT-4( ) are sealed and require no lubrication.

NEAR SURFACE REFERENCE TEMPERATURE DEVICE (NSRT)

The NSRT system consists of a meter assembly and a thermistor probe used to obtain

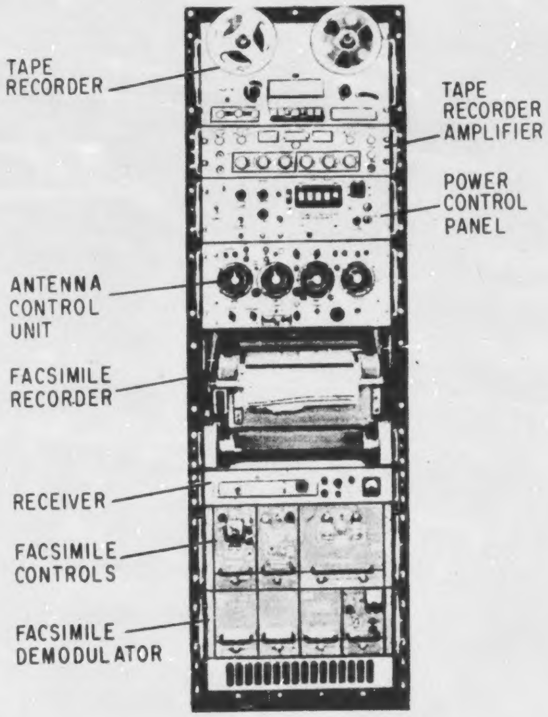


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Figure 18-17.–Receiver Recorder Set, Meteorological Data, AN/SMQ-6(V).

instantaneous measurements of sea surface temperatures. A description of this instrument along with a discussion of operating procedures may be found in AG 3 & 2, NT 10363-D.

MAINTENANCE

The NSRT system requires little or no maintenance other than occasionally replacing the flashlight battery. If maintenance is required, the IC electrician does the required work.

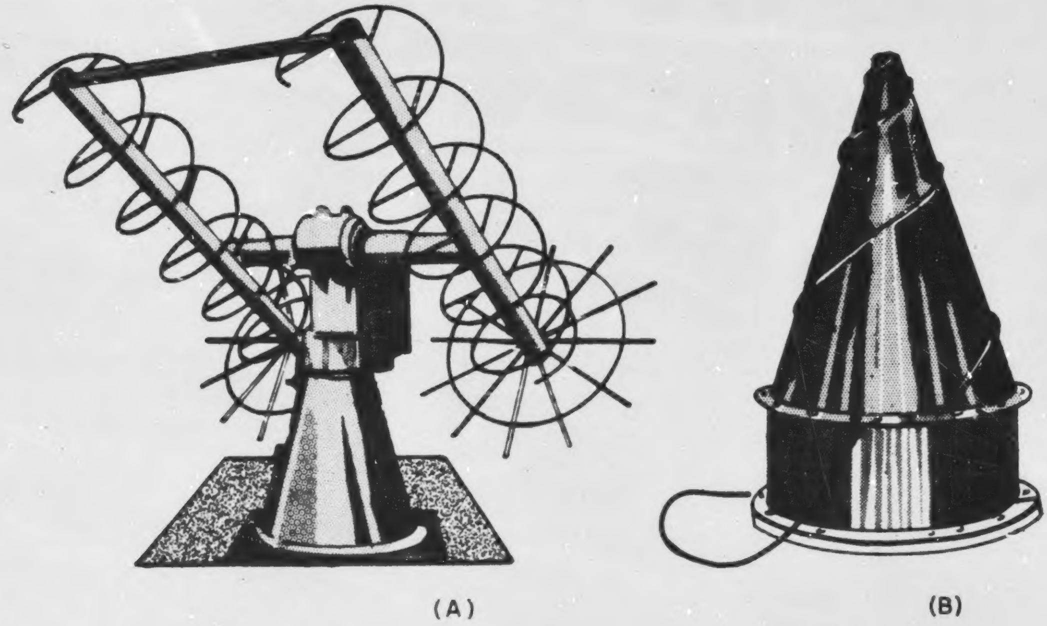
CALIBRATION

To calibrate the indicator, adjust the needle to the red line on the face of the meter (replace the battery if the needle cannot be adjusted to the red line). The instrument should be calibrated daily following this procedure.

SATELLITE GROUND RECEIVING EQUIPMENT

AUTOMATIC PICTURE TRANSMISSION (APT) RECEIVING AND RECORDING EQUIPMENT

Automatic picture transmission via weather satellite has become a common means of obtaining environmental data related to meteorological and oceanographic forecasting. The senior AG must insure that the receiving and recording equipment is kept in an operational status insofar as his maintenance responsibility extends. Electronics maintenance personnel will be responsible for most of the necessary tests, calibration, and maintenance; however, the AG may perform certain maintenance related tasks as described in the following paragraphs.



AG.183

Figure 18-18.—(A) Antenna, Dual Array Helical AS2192/SMQ-6(V); (B) Antenna, Fixes Omnidirectional, Conical AS2191/SMQ-6(V).

MAINTENANCE

There are a number of APT receiver recorders in use in the field as discussed in the AG 3 & 2 Rate Training Manual. Only the maintenance tasks pertaining to Receiver Recorder Set AN/SMQ-6(V) are presented here due to the similarity of the maintenance requirements of the various types of equipment. Figure 18-17 illustrates Receiver Recorder Set AN/SMQ-6(V). For further information pertaining to other equipment, refer to the applicable technical manual.

Figure 18-18 illustrates the two types of antennas which may be utilized. If the fixed omnidirectional antenna is used, the antenna control panel on the receiver recorder is replaced by a blank panel since it is not required.

The maintenance which will be performed on this equipment by the Aerographer's Mate operator is contained in NAVAIR 50-30GMH6-1.

CHAPTER 19

RADIOSONDE AND RAWINSONDE EQUIPMENT MAINTENANCE AND CALIBRATION

The upper-air observation program is one of the more important of the various programs required by the environmental services. Quality upper-air data is vitally important to the forecaster in the preparation of required forecasts. The prerequisite for collection of accurate upper-air data is the use of properly maintained, tested, and calibrated radiosonde/rawinsonde equipments.

The maintenance, test, and calibration procedures presented in this chapter are at the AG 1 & C level. For details on the component parts and operation of the equipments covered, refer to the applicable equipment manual.

RADIOSONDE RECEPTOR AN/SMQ-1( )

Radiosonde Receptor AN/SMQ-1( ) is a radio receiving and recording device operating in the frequency range of 390 to 410 mHz. It is used by the Aerographer's Mate to receive and graphically record meteorological data transmitted from a balloon-borne radiosonde transmitter (AN/AMT-11( )), by means of an RF carrier pulse modulated at audio rates which correspond to the ambient temperature and humidity of the air through which the radiosonde passes.

The receptor is designed for maximum reliability and life under adverse conditions of shock, vibration, inclination, temperature, and humidity encountered in shipboard use. To a great degree, it is self-compensating for wear, aging, and fluctuations in line voltage. When properly mounted, installed, and alined in accordance with the NavAir technical manual, it accurately records transmitted data using operating controls only. (See fig. 19-1.)

Many delicate pieces of electronics equipment require that an adequate preventive maintenance program be implemented to prevent unwarranted breakdowns and interruptions to the operational program for which they are designed. The Radiosonde Receptor is no exception to this rule. This equipment must also be calibrated monthly and whenever a technician replaces electronic parts or makes any mechanical adjustments to the pen drive system to insure that it is operating within certain limits of acceptable accuracy.

INSTALLATION

The location and the mounting of the AN/SMQ-1 are discussed in the technical manual for the equipment.

The antenna normally used with this equipment is the folded ground plane type. The applicable instructions manual should be followed when installing the antenna assembly.

MAINTENANCE

Maintenance Requirements Cards (MRC's) have been prepared for use by maintenance personnel when performing maintenance tasks on the AN/SMQ-1( ). (See fig. 19-2.)

The operator's maintenance on the AN/SMQ-1( ) is normally limited to routine cleaning and minor mechanical maintenance. All other maintenance should be performed by a qualified electronics technician.

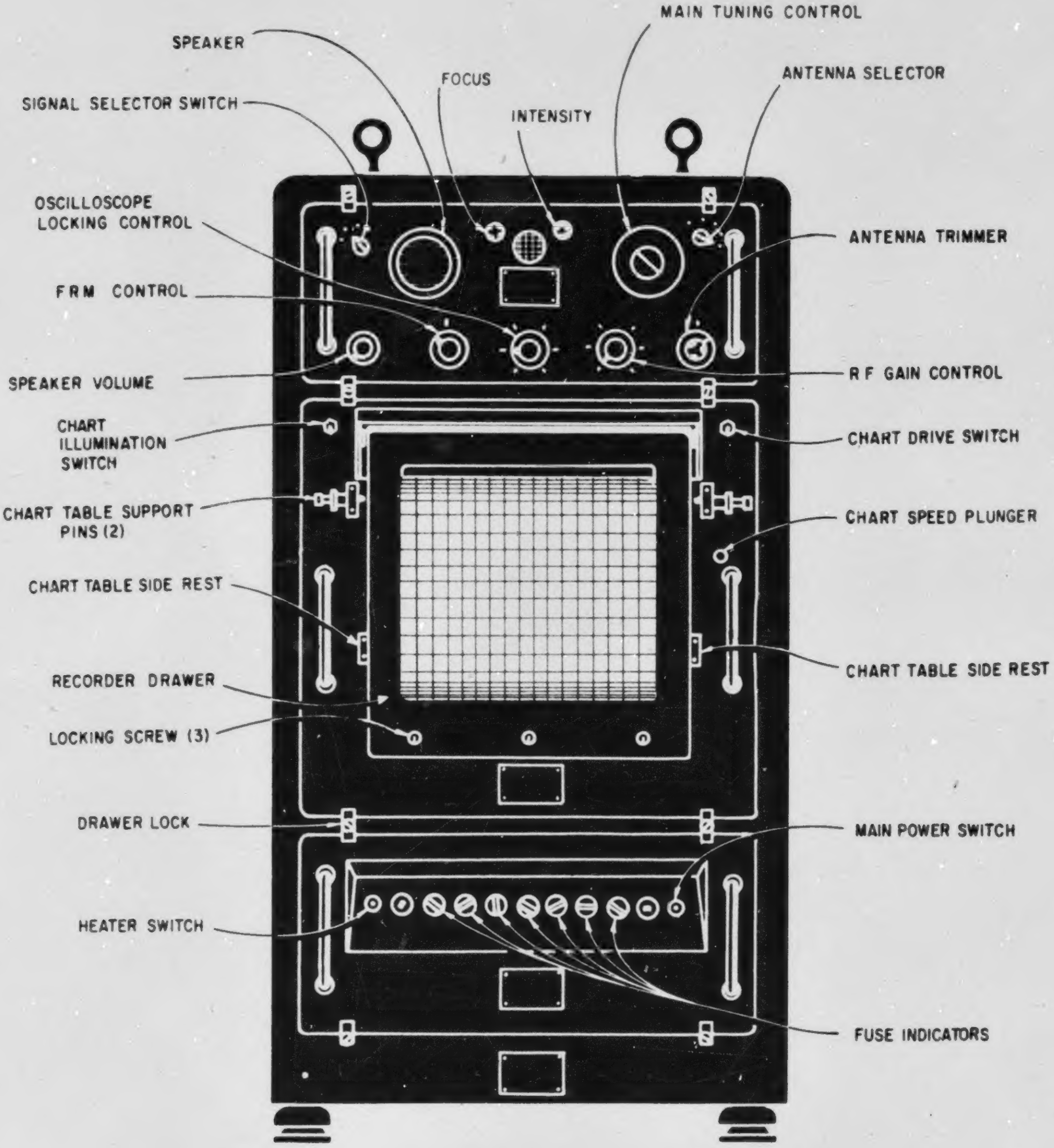
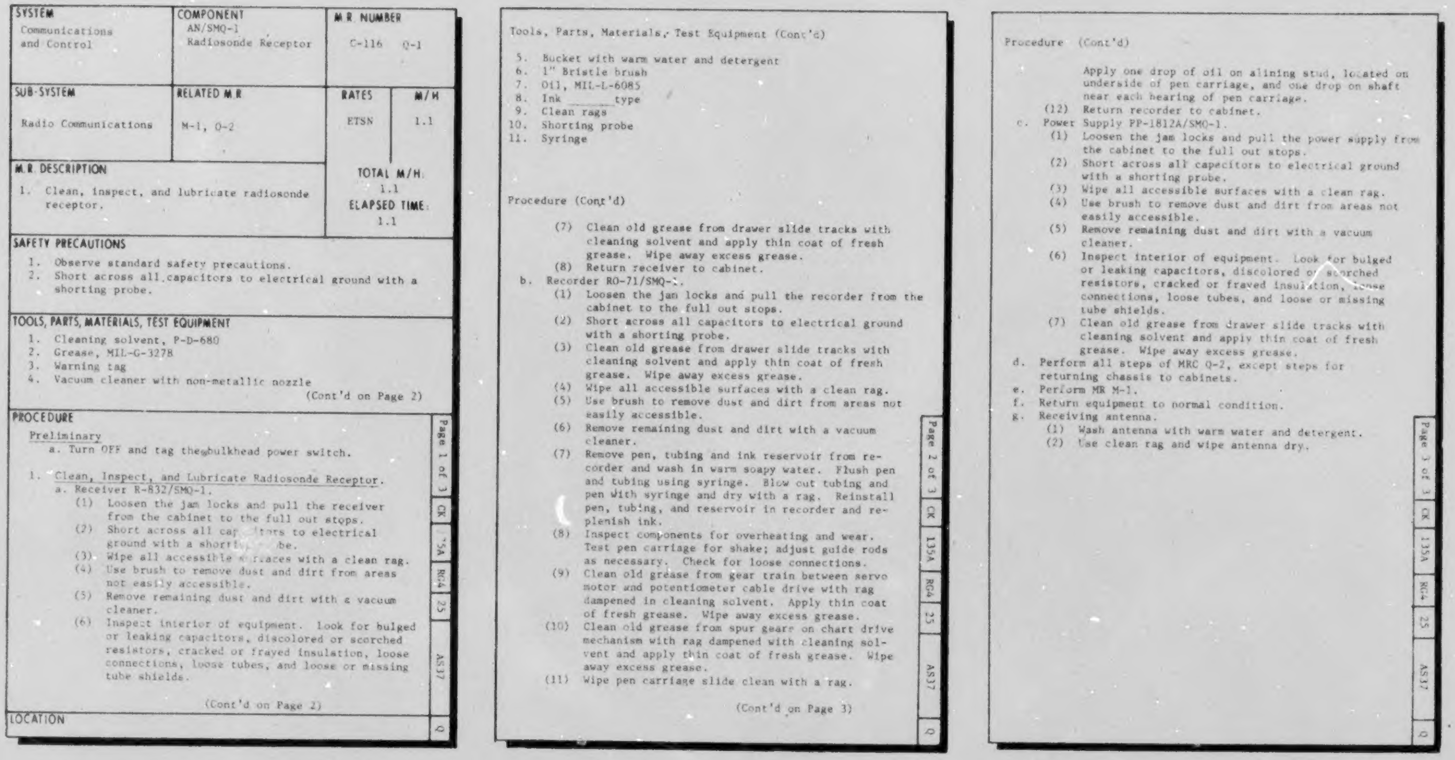

MAIN TUNING CONTROL
SPEAKER
FOCUS
ANTENNA SELECTOR
SIGNAL SELECTOR SWITCH
INTENSITY
OSCILLOSCOPE LOCKING CONTROL
ANTENNA TRIMMER
FRM CONTROL
SPEAKER VOLUME
R F GAIN CONTROL
CHART ILLUMINATION SWITCH
CHART DRIVE SWITCH
CHART TABLE SUPPORT PINS (2)
CHART SPEED PLUNGER
CHART TABLE SIDE REST
CHART TABLE SIDE REST
RECORDER DRAWER
LOCKING SCREW (3)
DRAWER LOCK
MAIN POWER SWITCH
HEATER SWITCH
FUSE INDICATORS
AG.751

Figure 19-1.—Radiosonde Receptor AN/SMQ-1( ).



SYSTEM	COMPONENT	M.R. NUMBER	
Communications and Control	AN/SMQ-1 Radiosonde Receptor	C-116 Q-1	
SUB-SYSTEM	**RELATED M.R.**	**RATES**	**M/H**
Radio Communications	M-1, Q-2	ETSN	1.1
M.R. DESCRIPTION		**TOTAL M/H:** 1.1 **ELAPSED TIME:** 1.1	
1. Clean, inspect, and lubricate radiosonde receptor.			

SAFETY PRECAUTIONS

1. Observe standard safety precautions.
2. Short across all capacitors to electrical ground with a shorting probe.

TOOLS, PARTS, MATERIALS, TEST EQUIPMENT

1. Cleaning solvent, P-D-680
2. Grease, MIL-G-3278
3. Warning tag
4. Vacuum cleaner with non-metallic nozzle

(Cont'd on Page 2)

PROCEDURE

Preliminary
a. Turn OFF and tag the bulkhead power switch.

1. Clean, Inspect, and Lubricate Radiosonde Receptor.
a. Receiver R-832/SMQ-1.
(1) Loosen the jam locks and pull the receiver from the cabinet to the full out stops.
(2) Short across all capacitors to electrical ground with a shorting probe.
(3) Wipe all accessible surfaces with a clean rag.
(4) Use brush to remove dust and dirt from areas not easily accessible.
(5) Remove remaining dust and dirt with a vacuum cleaner.
(6) Inspect interior of equipment. Look for bulged or leaking capacitors, discolored or scorched resistors, cracked or frayed insulation, loose connections, loose tubes, and loose or missing tube shields.

(Cont'd on Page 2)

LOCATION

Page 1 of 3	CK	5A	RG4	25	AS37	Q

Tools, Parts, Materials, Test Equipment (Cont'd)

5. Bucket with warm water and detergent
6. 1" Bristle brush
7. Oil, MIL-L-6085
8. Ink ________ type
9. Clean rags
10. Shorting probe
11. Syringe

Procedure (Cont'd)

(7) Clean old grease from drawer slide tracks with cleaning solvent and apply thin coat of fresh grease. Wipe away excess grease.
(8) Return receiver to cabinet.
b. Recorder RO-71/SMQ-1.
(1) Loosen the jam locks and pull the recorder from the cabinet to the full out stops.
(2) Short across all capacitors to electrical ground with a shorting probe.
(3) Clean old grease from drawer slide tracks with cleaning solvent and apply thin coat of fresh grease. Wipe away excess grease.
(4) Wipe all accessible surfaces with a clean rag.
(5) Use brush to remove dust and dirt from areas not easily accessible.
(6) Remove remaining dust and dirt with a vacuum cleaner.
(7) Remove pen, tubing and ink reservoir from recorder and wash in warm soapy water. Flush pen and tubing using syringe. Blow out tubing and pen with syringe and dry with a rag. Reinstall pen, tubing, and reservoir in recorder and replenish ink.
(8) Inspect components for overheating and wear. Test pen carriage for shake; adjust guide rods as necessary. Check for loose connections.
(9) Clean old grease from gear train between servo motor and potentiometer cable drive with rag dampened in cleaning solvent. Apply thin coat of fresh grease. Wipe away excess grease.
(10) Clean old grease from spur gears on chart drive mechanism with rag dampened with cleaning solvent and apply thin coat of fresh grease. Wipe away excess grease.
(11) Wipe pen carriage slide clean with a rag.

(Cont'd on Page 3)

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Procedure (Cont'd)

Apply one drop of oil on alining stud, located on underside of pen carriage, and one drop on shaft near each bearing of pen carriage.
(12) Return recorder to cabinet.
c. Power Supply PP-1812A/SMQ-1.
(1) Loosen the jam locks and pull the power supply from the cabinet to the full out stops.
(2) Short across all capecitors to electrical ground with a shorting probe.
(3) Wipe all accessible surfaces with a clean rag.
(4) Use brush to remove dust and dirt from areas not easily accessible.
(5) Remove remaining dust and dirt with a vacuum cleaner.
(6) Inspect interior of equipment. Look for bulged or leaking capacitors, discolored or scorched resistors, cracked or frayed insulation, loose connections, loose tubes, and loose or missing tube shields.
(7) Clean old grease from drawer slide tracks with cleaning solvent and apply thin coat of fresh grease. Wipe away excess grease.
d. Perform all steps of MRC Q-2, except steps for returning chassis to cabinets.
e. Perform MR M-1.
f. Return equipment to normal condition.
g. Receiving antenna.
(1) Wash antenna with warm water and detergent.
(2) Use clean rag and wipe antenna dry.

Page 3 of 3	CK	135A	RG4	25	AS37	Q

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Figure 19-2.—Maintenance Requirement Card for AN/SMQ-1( )

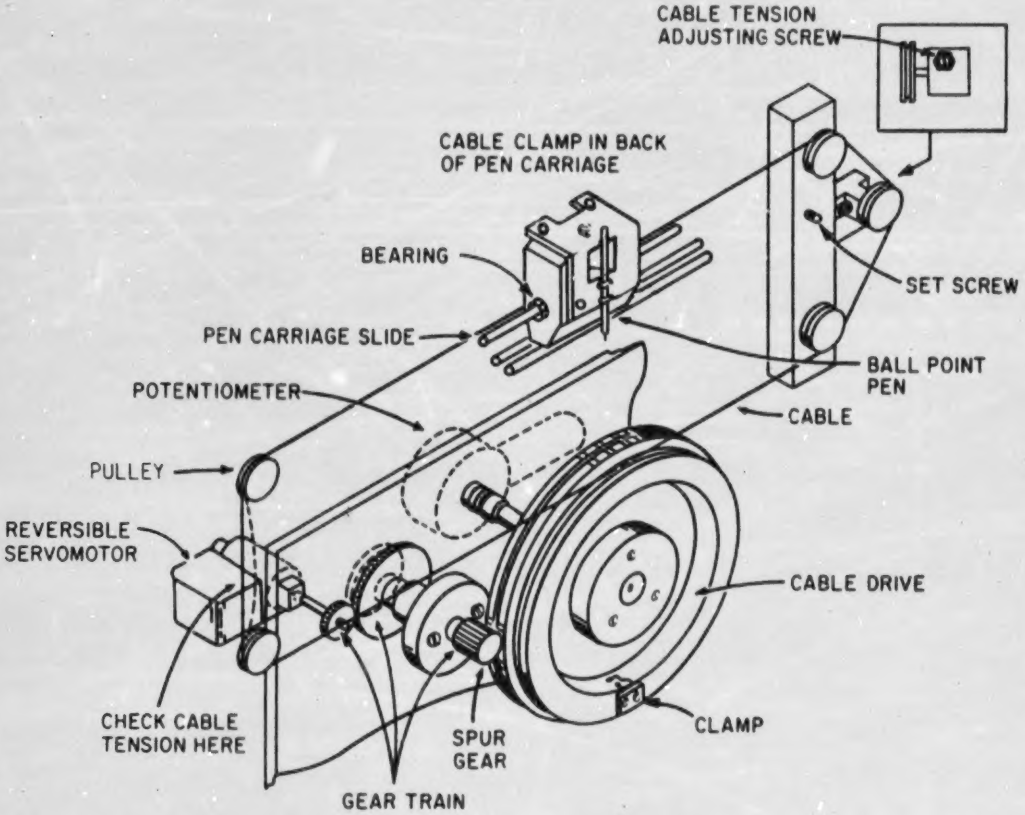
CABLE TENSION ADJUSTING SCREW
CABLE CLAMP IN BACK OF PEN CARRIAGE
BEARING
SET SCREW
PEN CARRIAGE SLIDE
BALL POINT PEN
POTENTIOMETER
CABLE
PULLEY
REVERSIBLE SERVOMOTOR
CABLE DRIVE
CLAMP
CHECK CABLE TENSION HERE
SPUR GEAR
GEAR TRAIN

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Figure 19-3.–Pen drive system, AN/SMQ-1( ).

Cleaning

The operator should clean the exterior and inside work area daily with a clean, lint-free cloth. Check the legibility of the ball point pen trace daily. If the trace is not clear and continuous, wipe the penpoint with a clean, lint-free cloth. If the discrepancy is not corrected by this procedure, replace the ballpoint pen.

Mechanical Maintenance

Mechanical maintenance of the Radiosonde Receptor AN/SMQ-1( ) consists of certain inspections, tests, checks, and adjustments which can be performed on an operator's level.

DAILY MAINTENANCE.–Perform the following tests and inspections daily, and adjustments as required:

1. Inspect the pen drive system cable for fraying (fig. 19-3).
2. Test the cable tension and adjust if necessary. Cable tension is tested on the left-hand side of the recorder drawer between the two pulleys (fig. 19-3). The cable should deflect 1/16 to 3/16 inch under slight finger pressure. The pen cable adjustment is a screwdriver adjustment found on a sliding block to which is attached an idler pulley located on the right side of the recorder drawer. A setscrew which holds the sliding block may have to be loosened to relax the cable tension, but simply turning the screw

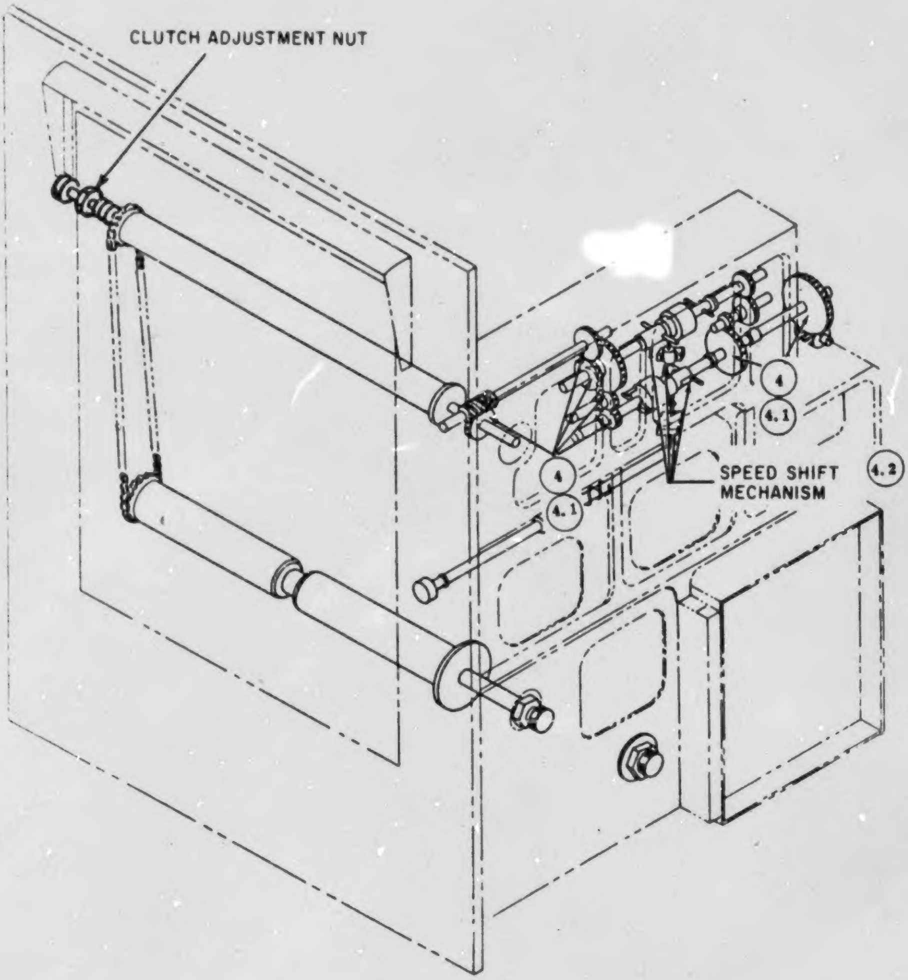

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Figure 19-4.—Chart drive system, AN/SMQ-1( ).

clockwise will tighten the cable tension. If repeated adjusting causes the cable to stretch to the point where a fully clockwise adjustment has no effect, a new cable will have to be installed.

3. Inspect the cable clamps and tighten if necessary (fig. 19-3).

4. Test the tension of the paper takeup clutch on the drive roll. If the paper is loose on the takeup roll or the holes are torn on the side of the direction of feed, there is insufficient tension. The clutch is tightened by turning a flat nut on the drive roll on the left-hand side of the recorder drawer. Turn the nut to the right onehalf turn at a time (fig. 19-4).

WEEKLY MAINTENANCE.—Perform the following checks, tests, and adjustments weekly if necessary:

1. Turn the chart drive ON, and using a stopwatch or sweep second-hand, measure the paper feed during a 10-minute run. If the paper feeds out less than 4 13/16 inches in 10 minutes, tighten the takeup clutch as indicated in the

daily adjustment. If it feeds more than 5 3/16 inches, have the technician check the line frequency.

2. Test the pen carriage for shake and looseness. Adjust guide rods if necessary (fig. 19-3).

3. Inspect the speed shift mechanism on the paper drive and tighten setscrews if necessary (fig. 19-4).

SEMIANNUALLY. Perform the following inspections semiannually and corrective action as necessary:

1. Inspect the cabling. Examine carefully all the interconnecting cables, particularly those between the drawers and the cabinet for evidence of chafing or damage. Have the technicians repair or replace as necessary.

2. Inspect the antennas. Examine carefully all the antenna leads and antenna for mechanical damage or corrosion. Have the technicians replace or repair as necessary.

3. Inspect the mounts. Inspect the shock mounts and the standoffs. Replace the damaged mounts. Tighten all mounting bolts.

TROUBLESHOOTING

Troubleshooting of the Radiosonde Receptor AN/SMQ-1( ) is of an electronics nature and is performed by the electronics technicians in accordance with the applicable NavAir technical manual.

CALIBRATION OF RADIOSONDE RECEPTOR AN/SMQ-1( )

The primary purpose of calibration of equipment is to determine any corrections that are necessary to make the actual operation of the equipment the same as the theoretical operation. In other words, the theoretical operation of the equipment is such that with a given input frequency the pen will move precisely to a predetermined ordinate value. Calibration provides a convenient method of checking to determine whether the equipment is operating within certain limits of accuracy. If the equipment does not operate within these limits, it is necessary to procure the services of an electronics technician to realine the circuits so it will operate properly.

Radiosonde Receptor AN/SMQ-1( ) should be calibrated once each month between the last sounding of the month and the first sounding of the subsequent month (or as soon thereafter as possible), and whenever equipmental repairs and/or adjustments have in any way affected the the previous calibration. Signal Generator SG-21B/U supplies the audio input frequencies which are necessary if accurate computation of frequencies from the printed cords are to be made. The original and copies of these records are forwarded to appropriate agencies monthly.

Signal Generator SG-21B/U

The Signal Generator SG-21B/U is a portable precision piece of calibration test equipment that is designed to produce an audio signal between 10 and 500 cycles per second with an accuracy of plus or minus 0.05 percent of the indicated frequency.

It provides a low audiofrequency signal with which to calibrate the AN/SMQ-1( ) and the AN/TMQ-5( ) Weather Data Recording Sets. For specific instructions regarding the steps to follow during calibration of the AN/SMQ-1( ) and the AN/TMQ-5( ) refer to the applicable technical manual.

Computing Recorder Correction Curves, Graphs, and Forms

The following steps outline the procedure for computing the recorder correction curves, graphs, and associated forms:

1. Read the ordinate values of the printings on the recorder calibration record to the nearest 0.1 ordinate and record them in the appropriate columns of Data Sheet No. 1 (fig. 19-5). Be sure and enter the figures for the "down" run in the DOWN column and the "up" run in the UP column.

2. Compute the mean value to the nearest 0.05 ordinate.

3. Compute the final corrections to the nearest 0.1 of an ordinate and enter them in the appropriate column of Data Sheet No. 1. Affix the proper sign to the correction. The correction

DATA SHEET NUMBER ONE

INPUT FREQUENCY CYCLES	SHOULD READ	RECORDER PRINTING READS (SCALE SWITCH IN 250 POSITION)			RECORDER AVERAGE CORRECTION
		DOWN	UP	MEAN	
190	95.0	95.0	95.0	95.0	0
170	85.0	84.9	84.9	84.9	+.1
150	75.0	74.9	74.9	74.9	+.1
130	65.0	65.0	64.9	64.95	+.1
110	55.0	55.0	55.0	55.0	0
90	45.0	44.9	45.0	44.95	+.1
70	35.0	35.0	34.9	34.95	+.1
50	25.0	24.9	24.9	24.9	+.1
30	15.0	15.0	14.9	14.95	+.1
10	5.0	5.0	5.0	5.0	0
GND					

RECEPTOR SERIAL NUMBER 28

SIGNAL GENERATOR SERIAL NUMBER 128

DATE 1 AUGUST 1972

OPERATOR AG2 H. J. Mathews USN

SHIP OR STATION USS ROOSEVELT (CVA 42)

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Figure 19-5.—Data Sheet No. 1, AN/SMQ-1( ) calibration.

should be such that when applied to the MEAN column it will correct that reading to read the same as the SHOULD READ column.

4. Using the correction on the data sheet versus the ordinate value on the recorder chart, construct a graph of the recorder corrections on the recorder record in the space provided by the 1 foot above the calibration run. (See fig. 19-6, upper half.)

This procedure is as follows:

a. Select one of the horizontal printed lines to be the "0" (ZERO) correction. Label this line accordingly.

b. Above the zero correction line, label each horizontal line with +0.1, +0.2, +0.3, etc., and below the zero correction line label each horizontal line with -0.1, -0.2, -0.3, etc.

c. Plot each RECORDER AVERAGE CORRECTION point, using the ordinate value and correction value as arguments on the graph. Label the graph, RECORDER CORRECTION GRAPH. (See fig. 19-6.)

5. Compute a recorder correction table from the graph drawn on the recorder record. The table is entered above the graph on the left side of the chart and neatly labeled, RECORDER CORRECTION TABLE. (See fig. 19-6.) To compute the values for this table, proceed as follows:

a. Start at 95.0 and move to the left on the graph until the graph line departs from zero by 0.05 correction and note the ordinate value at this point. On the recorder correction table, enter 95.0 to ____ equals 0.0.

b. At the last ordinate value selected, move to the left until the graph line departs from that correction by 0.05. Enter the data in the same manner as in a. above.

c. Follow this procedure until all the corrections have been entered down to the 5.0 ordinate value.

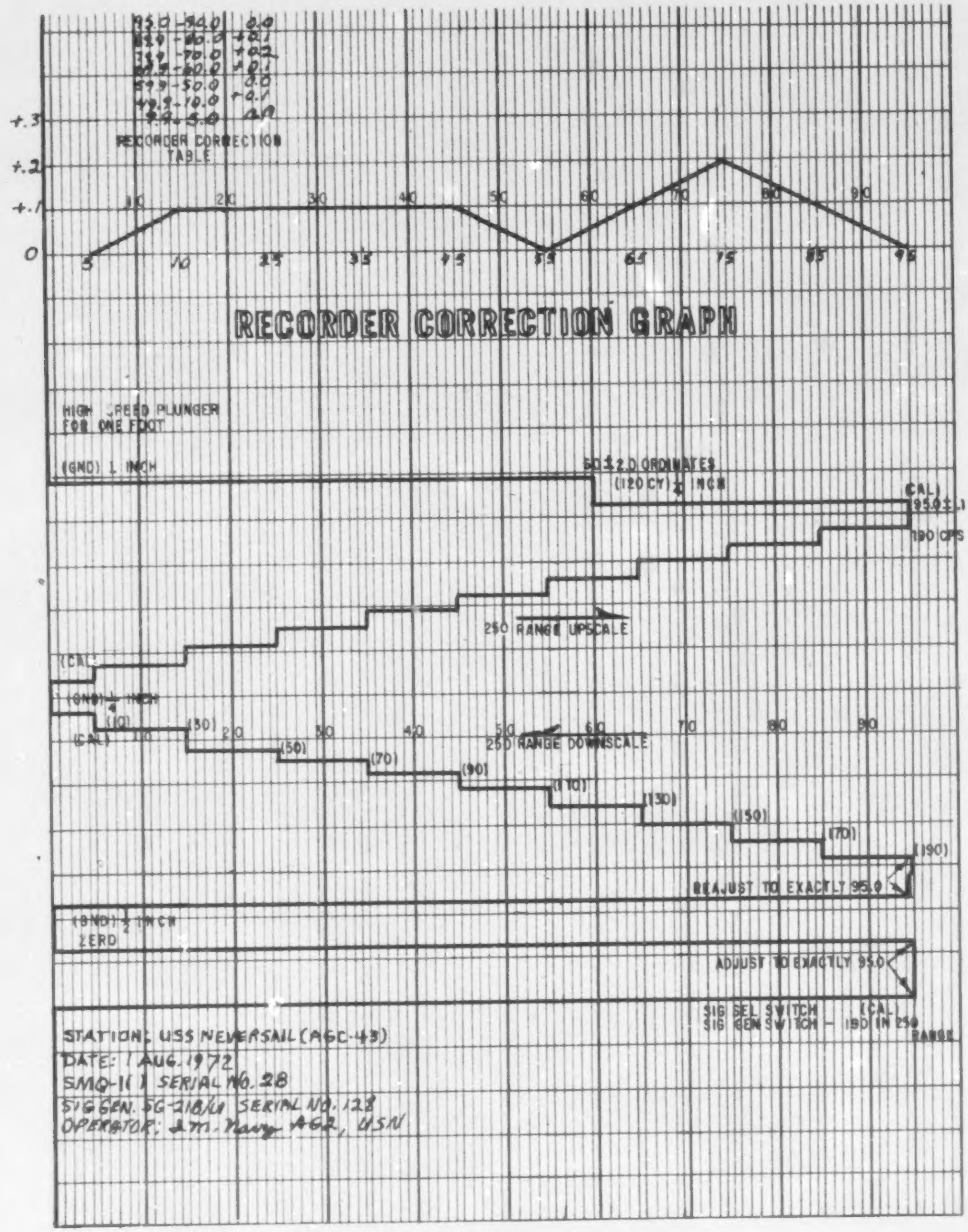
6. Make duplicate copies of the calibration correction and attach them to the recorder in a convenient location for the operator to apply during an observation.

Completing Calibration Record and Submission of Reports

After completing the calibration run, fold, and enter the following information as illustrated in figure 19-6 in the lower 7 inches of the chart:

1. Station identification and geographical location.
2. Date of calibration.
3. Serial number of the radiosonde receptor.
4. Serial number of the signal generator used.
5. Initials, name, and rate of the operator who calibrated the equipment.

After entering this information, fold the entire record into 7-inch accordian folds so that the lower 7 inches will face outward.



95.0 - 94.0	0.0
93.9 - 80.0	+0.1
79.9 - 70.0	+0.2
69.9 - 60.0	+0.1
59.9 - 50.0	0.0
49.9 - 10.0	+0.1
9.9 - 5.0	0.0

RECORDER CORRECTION TABLE

STATION: USS NEVERSAIL (AGC-43)
DATE: 1 AUG. 1972
SMQ-1( ) SERIAL NO. 28
SIG GEN. SG-218/U SERIAL NO. 128
OPERATOR: I.M. Navy AG2, USN

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Figure 19-6.—Sample Calibration Record AN/SMQ-1( ).

Prepare the data sheet in typewritten original and one copy. Forward original of each calibration, affixed to the first sounding following that calibration, to NWRC when forwarding the monthly records. Include a list of the new recorder calibration corrections on the first 7-inch fold of the recorder record for the first sounding after the new calibration.

The address of NWRC is as follows:

Officer in Charge
NWSED
Asheville, N.C. 28801

NOTE: The calibration recorder record is mailed with the data sheet.

The copy is retained on file by weather office making calibration.

NOTE: It is very important that the zero setting of each observation agrees with the zero setting of the latest calibration. The calibrator must make certain that a sample of the calibration zero tracing, with a list of recorder corrections, is posted in full view of the operator.

RECEIVING SET, RADIOSONDE AN/SMQ-3

The AN/SMQ-3 is similar in appearance, has similar controls, and performs the same function as the AN/SMQ-1( ) receptor. However, there are some significant differences between these two receptors and the operator should be aware of them. These differences are as follows:

1. The AN/SMQ-3 has a selector switch that provides for a chart speed of 1, 2, or 4 inches per minute.
2. The signal selector switch has an RS-1 and an RS-2 position. The RS-1 position is used for normal AN/SMQ-3 type sounding. The RS-2 position is used at land stations when the receptor is connected to a GMD control recorder and is being used in place of a TMQ-5 recorder.
3. The signal selector must be in the RS-1 or -2 position before an image will appear in the oscilloscope. In other words, the signal cannot be tuned in while the selector is in the GMD position.

Equipment maintenance is performed in the same manner as was discussed earlier in this chapter for the AN/SMQ-1( ).

HUMIDITY CHAMBER ML-428/UM

Operating instructions for the humidity chamber are contained in the applicable NavAir technical manual.

Maintenance on the AG level consists mainly of periodically cleaning the inside of the chamber and covering the internal walls with paste wax to prevent the chamber from absorbing moisture. Also check and inspect for rust on internal metal parts and take preventive action as required. The motor blower should be oiled at both bearing points after each 4 hours of continuous operation, or weekly as required. If excessive overheating of the motor occurs, check the shaft clearance to insure that movement is free. The test switch motor needs no oiling.

BATTERY TEST SET AN/AMM-1( )

Battery Test Set AN/AMM-1( ) has been designed to test the BA-353/AM battery and the radiosonde set AN/AMT-11( ) before each sounding, in order to insure a high percentage of successful flights. The test set is a portable instrument which may be hand-carried. The entire unit, including cables, is housed in a sealed metal case. The front panel contains two meters, a circuit selector switch, four push-button switches, and a power adjustment panel.

Since this test set is not in widespread use, it is covered only briefly here. For further information on this test set see the applicable NavAir technical manual.

INSPECTION

The frequency of periodic inspections of the test set is determined by the amount of use it receives and by the manner in which it is handled. With normal daily use and with reasonable care in handling, a weekly check should suffice in insuring that it remains in a good operating condition. It is recommended that the cover be closed when the test set is not being used and that it be kept as free as possible from dust, dirt, grease, and salt water spray.

MAINTENANCE

The test set is designed to require a relatively small amount of maintenance. There are no parts requiring lubrication, and maintaining the meter calibration is the major service function.

RAWINSONDE SYSTEMS MAINTENANCE

It has been stated previously that maintenance of electronic components of meteorological equipment is to be accomplished through the utilization of trained electronics personnel. However, the Aerographer's Mate must evaluate each instance to determine if the necessity of notifying electronics personel exists. If the malfunction is determined to be due to some minor cause which does not require a technician, it may be remedied by operating personnel. Various tasks related to preventive maintenance should also be performed by operating personnel. In each instance the experience of the operating personnel will be an important factor. Generally speaking, if it is necessary to work internally on the equipment or if the use of meters and other test equipment are required, a technician must be called. Keep in mind that the continued operation of partially defective equipment may result in more serious and expensive damage to related components. The following paragraphs are included not to encourage the AG to attempt maintenance beyond his training; but to allow him some measure of knowledge in determining causative factors pertaining to malfunctions of the Rawinsonde Systems.

RAWIN SET AN/GMD-2( ) MAINTENANCE PROCEDURES

A well-organized preventive maintenance program will help keep equipment failures at a minimum level, but cannot prevent all failures. How long your equipment stays inoperative after a failure generally is governed by two factors: how quickly you can determine which component is defective, and how quickly you can either correct the trouble or have a spare component installed in place of the defective one. The length of time it takes to locate and correct a trouble is generally measured by how well you know the equipment, and how quickly you can spot unusual conditions and analyze them.

Often, trouble is caused by a defective tube, fuse, or other easily replaceable parts, and can be corrected simply by notifying electronics personnel for replacement of available parts. However, if troubleshooting and repair will take time which is needed for operational use of the equipment, or if realinement will be needed after repair, technicians will probably replace the entire component. Some components (such as power supplies, etc.) can be replaced without alinement. Other components might take technicians longer to aline into the system than the time it would take them to replace the defective components.

A good organizational/field maintenance operation is one in which electronics personnel keep all allowable spare components ready to fit into the system quickly with a minimum of alinement time needed. During non-flight periods, spare components are normally checked by the technicians and the necessary alinements and adjustments made so that they can be installed in the equipment with little or no delay.

PREVENTIVE MAINTENANCE

Routine maintenance procedures that will help the AG prevent troubles which would interfere with the performance of the equipment are contained in the set of Preventive Maintenance Workcards, NavAir 50-30GMD2-4. These routines include system performance checks and inspection, cleaning, and lubrication procedures which, if performed at regular intervals by the AG or technicians as required, will insure against premature failure or breakdown of the equipment components.

CORRECTIVE MAINTENANCE

Corrective maintenance consisting of preflight checks and the system performance test procedures described in the technical manual will enable the responsible parties to detect troubles in the equipment effectively. Troubles generally will be attributed to fuse failure or to defective electron tubes or wiring. Most other troubles can be corrected by the technician accomplishing proper alinement or adjustment. The alinement routine workcards for the equipment contain complete, detailed alinement instructions.

DEPOT LEVEL MAINTENANCE

Depot level maintenance may be required for maintenance tasks which, normally, cannot be accomplished at organizational/field levels. Special test equipment or test fixtures are called for and/or no standard equipment will perform the required test functions. The maintenance of the AN/GMD-2( ) requiring depot level facilities is described in the equipment technical manual.

RAWIN SET AN/GMD-1( ) MAINTENANCE PROCEDURES

The maintenance procedures described for the AN/GMD-2( ) will also generally apply to the older AN/GMD-1( ) System. Leveling, orientation, weatherproofing, general lubrication instructions, and simple operator preventive maintenance of this system are covered in the technical manual for Rawin Set AN/GMD-1( ), NavWeps 16-30-GMD1-5.

PREVENTIVE MAINTENANCE OF AN/TMQ-5( )

General lubrication, corrosion removal, rustproofing, painting and other simple maintenance procedures are covered in the technical manual for Radiosonde Recorder AN/TMQ-5, NavWeps 50-30TMQ5-501.

Routine maintenance consists mainly of changing the recorder chart roll, filling the pen, keeping the instrument clean, and such corrective measures as set forth in the applicable technical manual. NOTE: When servicing the equipment, be extremely careful; high voltage is present. Turn the power off when changing the paper or the pen, and for all other functions not requiring the power to be on.

Calibration of the AN/TMQ-5( ) is performed by the operator and is performed in a manner similar to that discussed previously for the AN/SMQ-1( ). The applicable NavAir technical manuals contain details for performing this calibration.

PREVENTIVE MAINTENANCE OF AN/GMM-1( )

Preventive maintenance for the AN/GMM-1( ) which consists of daily, weekly, and monthly inspections and preventive maintenance are listed in the NavAir technical manual.

PREVENTIVE MAINTENANCE OF AN/GMM-3( )

Preventive maintenance procedures for the AN/GMM-3( ), which consist of daily, weekly, and monthly inspections, are listed in the NavAir technical manual.

CHAPTER 20

ADMINISTRATION, TRAINING, COMMUNICATIONS

Advancement to AG1 will generally elevate you to filling the position of section leader which will require you to spend more of your time in directing and controlling the work of others rather than performing it yourself. As an AG1 or AGC you will also become more deeply involved in the administrative functions of your unit or division.

The efficiency and effectiveness of any weather unit are directly related to how the administrative functions are performed. Under administrative functions we are referring to such areas as: proper indoctrination and orientation of new personnel; watch routines; leave and liberty policies; training; and submission of required records and reports.

It is also important that senior personnel be aware of the responsibilities and functions of the various elements under the Naval Weather Service Command, especially in regard to their unit's relationship with its controlling activity.

Senior Aerographer's Mates will need to possess an understanding of the various communications systems that are utilized within the weather units. Closely associated with communications is security. Senior personnel will be directly involved in the circulation, control, and possibly destruction of classified information.

In this chapter we will discuss these areas to indoctrinate personnel preparing for advancement, as well as prepare the proper foundation to enable you to carry out these assignments and duty in the most efficient manner when they are presented to you.

NAVAL WEATHER SERVICE COMMAND

The Naval Weather Service Command (NAVWEASERVCOM) was established by the Secretary of the Navy in July 1967. This was the result of upgrading of the then Naval Weather Service to a command status. The Commander Naval Weather Service Command (COMNAVWEASERV), who replaced the Director of the Naval Weather Service in this reorganization, reports directly to the Chief of Naval Operations.

MISSION

The mission of the Commander Naval Weather Service Command, promulgated by the Secretary of the Navy, is to command assigned activities of the Naval Weather Service Command; to insure the fulfillment of Department of the Navy meteorological requirements and Department of Defense requirements for oceanographic analysis/forecasts; and to provide technical guidance in meteorological matters throughout the naval services. Fleet Weather Centrals, Fleet Weather Facilities, and Naval Weather Service Facilities have concurrent responsibility to the Fleet Commander in Chief and the Chief of Naval Training who exercise authoritative control over the product output to meet requirements of the operating forces. The Headquarters, Naval Weather Service Command, provides staff support to the Commander in the fulfillment of his mission and functional responsibility.

ORGANIZATION

A number of changes within NAVWEASERVCOM have been instituted recently with more planned in the near future. The most sweeping change was the redesignation of selected Fleet Weather Facilities to Naval Weather Service Facilities (NAVWEASERVFAC). The establishment of a new NAVWEASERVFAC at Glenview is proposed to centralize Reserve functions. These organizations will fall under the Commander in the same category as weather centrals and remaining facilities. At this writing other anticipated changes are proposed to various NWSED's around the globe.

As finalization of plans and implementation takes place it is anticipated that COMNAVWEASER will distribute an up to date chart as a change to The Manual of the Naval Weather Service Command, NAVWEASERVCOMINST 5400.1 (Series).

FUNCTIONS

Commander Naval Weather Service Command has a number of varied functions to perform in support of his assigned mission. They are too lengthy for inclusion in this manual but they are found in entirety in the Manual of the Naval Weather Service Command, NAVWEASERVCOMINST 5400.1 (Series). It is recommended that you refer to this instruction for their listing.

NAVAL WEATHER SERVICE COMMAND ELEMENTS

In order to effectively carry out the assigned mission of the Naval Weather Service Command, the Commander has command and support responsibilities for designated field activites. In turn, these field activities have a number of elements, NWSED's, under their command.

FIELD ACTIVITIES

There are four types of field activities directly under COMNAVWEASERV: these are Headquarters, NAVWEASERVCOM; Fleet Numerical Weather Central; Fleet Weather Centrals and Facilities; and Naval Weather Service Facilities.

Headquarters, NAVWEASERVCOM is staffed by officers, enlisted, and civilian personnel who provide the Commander with the staff support required to perform his mission and functional responsibilities. Among his assistants located at headquarters are the Deputy Commander as well as assistants for Resources Management and Training, Command Operations, Plans and Policy, and Programs, Requirements and Scientific Services. The Senior Enlisted Advisor (SEA) to the Commander, Naval Weather Service Command is also a staff member.

The Fleet Numerical Weather Central coordinates the electronic computer effort of the NAVWEASERVCOM. Its multimission responsibilities include basic processing, NEDN support, centralized supply operations, BT Co-op program, satellite data handling center, and development and test of numerical techniques in meteorology and oceanography, among others.

Fleet Weather Centrals are assigned geographic areas of responsibility within which they perform the functions of data collection, centralized computer processing, and the dissemination of analyses, forecasts, warnings, and ship routing information. Fleet Weather Facilities provide assistance in the performance of these functions assigned to the weather centrals in their respective areas. In addition they are assigned specialized tasks such as hurricane forecasting, ASWEPS support, surf and swell forecasting, and ice reconnaissance coordination and forecasting.

In order to relieve centralized processing centers and Fleet Weather Centrals providing direct fleet support from the detailed duties of providing management and command for NWSED's, Naval Weather Service Facilities were recently established. These units will specialize in command and management of these detachments as well as function in other fields as directed by the Commander. Plans presently call for the realinement of NWSED's to fall under the command of one of the following Naval Weather Service Facilities: London, Pensacola, Jacksonville, or San Diego. In addition Pensacola will also function as headquarters relating to the technical phase of AG training and advancement.

AREA REPRESENTATIVES

The mission of the Naval Weather Service Command Representatives (NAVWEASERVCOM REPS) is assigned by COMNAVWEASERV. The approved mission is to represent

COMNAVWEASERV as directed and to provide liason with and assistance to Fleet Commanders in Chief and the Chief of Naval Training in the exercise of their authoritative control over the product output of Fleet Weather Centrals and Facilities.

These representatives report directly to COMNAVWEASERV. The following billet incumbents are ordered Additional Duty as NAVWEASERVCOM REP.

BILLET	TITLE
CINCPACFLT (STAFF) Fleet Meteorologist	NAVWEASERVCOM REP PAC
CINCLANTFLT (STAFF) Director Meteorology and Oceanography	NAVWEASERVCOM REP LANT
CO NAVWEASERVFAC LONDON	NAVWEASERVCOM REP EUR/MED
CO NAVWEASERVFAC PENSACOLA	NAVWEASERVCOM REP CNT

WATCH AND ROUTINE DUTIES

Most weather units operate 24 hours a day, 7 days a week. Units may operate a lesser number of hours but this will generally be in smaller units and will be dictated by the types of operations conducted or possibly by the hours of operation of the landing field. However, even at stations with restricted hours for aircraft operations, weather units may be required to operate around the clock to provide timely forecasts and warnings to local commands for the preservation of property and safety of personnel.

In order to have a smooth running system a watch list is published to show the unit's personnel the times that they will be required to work.

WATCH LIST

Watch lists are generally prepared by the senior Aerographer's Mate assigned to the unit and approved by the Meteorological Officer or the Officer-in-Charge/Chief Petty Officer-in-Charge, ashore.

There are many variations of watch lists and by the time you have advanced to the AG1 level you have undoubtedly gained much experience with them, and how they are used. Major consideration in perparing a watch list will of course be the number of personnel that may be utilized on the watch list. It is also important to consider when peak workloads occur to insure that an adequate number of personnel are available to handle the workload at that time.

In some units in which radiosonde or rawinsonde observations are made, it may be necessary to provide a separate watch list for personnel engaged in this work.

In all cases though, care should be taken in watch list preparation to equalize the number of hours worked and develop a smooth rotation of the various shifts.

Preparation of shipboard watch lists actually presents more of a problem due to the limited number of personnel, and the demands for personnel to assist in normal shipboard routine. The rotation of the hours of duty is not so easily achieved as ashore. One practice is to have each section stand a particular shift for the duration of a portion of a cruise. During in-port or stand-down periods the rotation of shifts can generally be accomplished in the easiest manner. Normally during in-port periods office routine is relaxed to a degree. When ships are in ports where shore weather units are located, the shore unit will normally provide required services such as charts and local area forecasts.

Table 20-1 provides an example of a typical watch list utilized at shore weather units. In this table section A averages 42 hours per week, sections B and C average 44 hours per week and section D averages 38 hours per week. If the time alloted for training is actually utilized as such, the average for each section will increase to 50 to 60 hours. Although average hours worked differ over this short period they will average equally over a longer period.

ROUTINE DUTIES

The primary function of the weather unit is to provide meteorological and oceanographic products and services to the naval service. To

Table 20-1.—Watch list—shore station.

Date	Day	00-08	08-16	16-24	In training	Liberty
1	TUE.........	A	B	C	....	D
2	WED.........	A	B	C	....	D
3	THU.........	A	B	C	D	....
4	FRI..........	D	A	B	....	C
5	SAT.........	D	A	B	....	C
6	SUN.........	D	A	B	....	C
7	MON.........	C	D	A	....	B
8	TUE.........	C	D	A	....	B
9	WED.........	C	D	A	B	....
10	THU.........	B	C	D	....	A
11	FRI..........	B	C	D	....	A
12	SAT..........	B	C	D	....	A
13	SUN..........	A	B	C	....	D
14	MON.........	A	B	C	....	D
15	TUE.........	A	B	C	D	....
16	WED.........	D	A	B	....	C
17	THU.........	D	A	B	....	C
18	FRI..........	D	A	B	C	....
19	SAT..........	C	D	A	....	B
20	SUN..........	C	D	A	....	B
21	MON.........	C	D	A	B	....
22	TUE..........	B	C	D	....	A
23	WED.........	B	C	D	....	A
24	THU..........	B	C	D	A	....
25	FRI..........	A	B	C	....	D
26	SAT..........	A	B	C	....	D
27	SUN..........	A	B	C	....	D
28	MON.........	D	A	B	....	C

Section A	Section B	Section C	Section D
1. AG1..........	1. AG1...........	1. AG1...........	1. AG1............
2. AG2..........	2. AG2...........	2. AG2...........	2. AG2............
3. AG3..........	3. AG3...........	3. AG3...........	3. AG3............
4. AG3..........	4. AG3...........	4. AGAN.........	4. AGAN..........
5. AGAN........	5. AGAN.........	5. AGAN.........	5. AGAN..........

SUBMITTED:
(signature)
Office Supervisor

APPROVED:
(signature)
Meteorological Officer

accomplish these tasks there are many routine duties to be performed by the Aerographer's Mates. A broad generalized breakdown of these duties is as follows:

1. Observe, record and report meteorological and oceanographic conditions at regular and specified intervals.
2. Code and send reports through regular communication channels to designated activities.
3. Receive and plot reports from other sources.
4. Maintain needed supplies.
5. Instrument maintenance.
6. Work up climatological data.
7. Provide any special environmental data requested by authorized users.

Along with these routine duties a number of jobs or duty assignments, of a nonmeteorological nature, will be assigned to weather unit personnel. This is more prevalent aboard ship than ashore. Some of these will include:

1. JOOD watches.
2. Shore patrol.
3. Departmental duties; i.e., security watches, duty petty officer, etc.
4. Working parties.
5. Damage control duties.

Although these duties seem small in number they will have to be seriously considered when preparing the watch list.

EMERGENCY BILLS

All weather units will have a Watch, Quarter, and Station Bill, which should be conspicuously posted and kept up to date. Although not an emergency bill in itself the W, Q, & S Bill provides information concerning the assignment of personnel during emergency conditions.

On shore stations this will generally include fire station assignments, shelter or duty assignment for destructive weather conditions, and team or shelter assignment for disaster control.

Fire stations within each weather unit will generally be assigned to personnel of the unit. Frequently Aerographer's Mates will be assigned to the disaster control party to act as observer or forecaster to work in conjunction with helicopter recovery operations. These personnel may be airlifted to remote landing sites and operate from there.

The shipboard Watch, Quarter, and Station Bill will encompass more areas. Its proper preparation will require reference to the Battle Organization Manual and the Ships Organization and Regulations Manual.

The W, Q, & S Bill includes the assignment of particular duties or stations to each person during the various emergency conditions, such as general quarters, fire, man overboard, collision, or nuclear, biological, and chemical defense.

As a senior petty officer you will be directly involved in the preparation of such bills. It is important that you familiarize yourself with all applicable instructions that govern them.

RECORDS AND REPORTS

Efficient adminsitration, direction, operation, and coordination of the Naval Weather Service necessitates information about the operation of each weather unit. Most of the information is obtained through the medium of regular reports and accurate records on the operation of each weather unit. Regularly recurring reports supply the Commander of the Naval Weather Service Command with this information.

Within this section we will discuss some of the required reports and records as well as the proper disposal methods.

REQUIRED CHARTS, RECORDS AND REPORTS

Weather units will determine on an individual basis the charts they will require to carry out their assigned duties. With the complete coverage afforded by the National Facsimile Network very few of the smaller units will locally produce charts. Larger units and those providing charts for Fleet Facsimile Broadcasts will employ locally produced and numerically produced charts. A number of National Facsimile charts will also be retransmitted over the Fleet Broadcast.

The Chief of Naval Operations requires all weather observing meteorological units to complete and submit monthly the Meteorological

Records Transmittal Form, NWSC 3140/6. Even during months when observations are temporarily discontinued, such as the ship being in the yard, the report is to be submitted. This form serves the function of a letter of transmittal for the monthly meteorological records. The instructions for filling out and mailing of the form are contained on the reverse side of the form.

The Meteorological Station Report Description and Instrumentation, NWSC Form 3140/5 will be prepared as of 1 January each year and submitted. It will also be prepared and submitted when there is a change which affects one or more of the entries on the form. This report will also include commissioning or decommissioning of a ship or station and periods of expansion or curtailment of meteorological activities. The original will be sent to NWSED Asheville with a copy to COMNAVWEASERV.

Naval Weather Service Command shore stations will also submit monthly a Report of Assigned Manpower/Personnel, NWSC 1080-4. Personnel status as of the last day of the month will be reflected on the report. This report should be submitted so that the original reaches Headquarters Naval Weather Service Command no later than the 10th of the following month.

In addition to these reports, units may have to submit other reports to their controlling activity. Such information as civilian work schedules, submission of time cards for pay purposes and budgetary information are some examples. These types of required reports will vary however.

DISPOSAL INSTRUCTIONS

The periods of retention and disposal methods of records and reports vary. Observational records will be submitted monthly for quality control checks at the National Climatic Center. Carbon copies of surface weather observations will be retained indefinitely ashore and for at least 2 years aboard ship. Locally prepared forecasts and warnings will be retained for 1 year, except for storm and small craft warnings, which are retained for 6 months.

For a detailed listing of applicable retention periods and disposal methods refer to SecNav Instruction 5215.5 (Series), Part II, Chapter 3.

TRAINING AND INSPECTION PROCEDURES

There are definite reasons why training is so important within the Naval service. Foremost among these is training of personnel to perform their assigned duties in the most effective and efficient manner. New equipment or methods require training in their proper operation or utilization, even to senior personnel. The preparation of personnel to participate in the advancement examinations also necessitates a thorough training program.

Proper operation and maintenance of equipment requires the formulating of proper inspection procedures. A comprehensive inspection program will in many cases, enable you to detect and prevent problems before they arise.

In this section we will discuss both of these areas in general.

TYPES OF INSTRUCTION

The desired results of training programs will dictate the manner in which training should be conducted within the individual unit. There are two general types of instruction: on-the-job type; and formal or classroom type of instruction. The most proficient type of training program will utilize each type of training to some degree.

Even though junior personnel may have completed school training in the rate, they will not be completely ready to perform all the varied tasks assigned them upon reporting to their duty station. The school has provided them with theory and basic information, but the on-the-job training in the individual unit will polish the skills that they will need to perform their tasks. This type of training will also familiarize personnel with the proper operation and maintenance of the various equipment they will be using.

On-the-job training is the most widely used type of training and there are reasons for this. It can normally be carried out during the normal work cycle of the individual and requires neither extra time nor personnel in conducting it. This type of training generally revolves around the section leader who will instruct the individual while he is performing the various tasks. The one-on-one relationship allows time and closeness of supervision so that the individual can normally learn at a rather rapid rate.

There are limitations to this type of training in that it can only train personnel in tasks that are being performed in that particular unit as well as allowing them to become familiar with equipment that the unit possesses. The effectiveness of the program is greatly dependent on the desire of the person conducting the training, as the trainee will learn only information and methods that his instructor imparts to him.

Some formal or classroom type of training should also be conducted. Here new equipment new methods or theoretical background can be taught. Also equipment not within the local area or operations not performed locally can be discussed. This type of training will generally be used in preparing personnel for future assignment to other units and advancement examinations and should be organized to include all data covered in the "Quals" Manual.

In both on-the-job and formal training, programed instruction can be used to advantage to provide individualized instruction.

In order to have an effective training program, whether it be on-the-job or formal classroom type, it is necessary to have a competent petty officer responsible for its operation. Training schedules should be prepared and adhered to. Records should be kept to substantiate the training each individual has had. Personnel assigned the responsibility of conducting the various segments of training should be notified well in advance so they may properly prepare their material.

Information pertaining to the duties of a petty officer as an instructor is covered in Military Requirements for Petty Officer 1 & C, NavTra 10057-C. The Manual for Navy Instructors, NavPers 16103-C, provides an excellent reference for personnel to utilize concerning the techniques of teaching that should be employed, as well as pertinent items such as lesson guides, lesson plans, and training aids.

One measurement of the effectiveness of a training program is normally considered to be the results of the advancement examinations. However, the overall proficiency and performance of any weather unit is dependent upon personnel being trained to do their job in the most efficient or effective manner. A proficient, smooth running organization will generally have a well organized and well run training program.

SCHEDULING INSPECTIONS

Most meteorological, oceanographic and NEDN tie-line equipment requires periodic inspection to insure its proper operation. This periodic inspection may be part of an outlined maintenance program or it may consist of visually inspecting the equipment only. Whichever it is, strict adherence to such procedure is a necessity.

As a senior petty officer, the responsibility for the formulation and administration of such procedures will be yours. This requires the organizing of records that show that the inspections are being conducted as scheduled. At the start you must make certain that personnel who are to conduct these inspections are familiar with the equipment and procedure. They should also be informed of some of the possible discrepancies they can expect to find. This may be accomplished by conducting training sessions on the particular equipment involved.

To make the inspection program do its job you must closely monitor it. Routine checking of both logs and the equipment itself will make your assigned personnel aware that you are interested in their doing a proper job. Failure to followup is probably the greatest pitfall of any inspection program. This allows for the "gundecking" of logs and reports without the inspections being performed.

The direct results of incorporating an effective inspection program within your unit is reduction in the amount of down time for each piece of equipment. A properly conducted and administered program will give maximum operational time on all equipment.

COMMUNICATIONS

Communications plays an important role in both the collection of weather data for evaluation and analysis and again in the dissemination of forecasts and warnings so they may be utilized to the fullest extent possible.

The most common type of communications used by weather personnel will be teletype and facsimile systems. Both of these systems have been thoroughly discussed in chapter 23 of AG 3 & 2, NavTra 10363-D.

In this section we will briefly discuss two other modes of communications you will come in contact with.

COMPUTER SYSTEMS

Computer systems are used within the Naval Weather Service to perform functions ranging from collection and processing of raw data; to preparation of the data for retransmission.

There are a number of different types of computer systems in use by the weather service. These computer systems are discussed in chapter 11 of AG 3 & 2, NavTra 10363-D.

The Navy Environmental Data Network employs computers in the distribution of its products.

NAVY ENVIRONMENTAL DATA NETWORK

This communications mode utilized within the weather service is more commonly referred to as NEDN.

Information transmitted via NEDN is received from two primary sources, the Automated Weather Network (AWN) of the U.S. Air Force and from stations connected to the NEDN.

The NEDN is used primarily for the collection of oceanographic and meteorological data from naval weather units, with oceanographic observations being of primary concern, and for retransmission of products to users. The NEDN consists of an interconnected system of digital computers and associated online telecommunications and peripheral equipment which has been integrated for the rapid collection, processing, dissemination and display of environmental data.

Additional information on the NEDN may be found in NAVWEASERVCOM INST 2309.1( ).

SECURITY

Everyone has the responsibility to safeguard classified information. As a senior petty officer you may expect to be assigned duties in the management of classified material as well as developing and supervising educational programs dealing with proper security procedures. Both of these duties are extremely important within any unit. In these sections we will discuss the various aspects of these duties that you should be cognizant of.

SECURITY CLASSIFICATION MANAGEMENT

Local management of classified material will be assigned by the Commanding Officer/Officer-in-Charge of the organization. The duties of the designated individual will cover all the aspects of the proper stowage, handling and destruction of classified material under his cognizance. He will be responsible for keeping of accurate records from receipt to destruction of classified material.

For complete information of duties as well as guidance in performance of assigned managerial duties reference should be made to OpNav Inst. 5510.1 (Series) and DOD Regulation 5200.1 (Series).

SECURITY EDUCATIONAL PROGRAMS

The security educational program of any organization should be designed to fit the particular requirements of personnel that have access to classified information. This program should include the following as a minimum:

1. Advise personnel of the need for protecting classified information and the adverse effects to the national security resulting from compromise.
2. Indoctrinate personnel fully in the principles, criteria, and procedures for the classification, downgrading, and declassification of material. Alert personnel to prohibition on improper use, and abuses of the classification and declassification exemption system.
3. Familiarize personnel with the specific security requirements of their particular assignment.
4. Make personnel aware of various techniques employed by foreign intelligence activities to obtain classified information, as well as the responsibility for reporting any attempts.
5. Advise personnel of the hazards and strict prohibitions on discussion of classified informa-

tion in any such manner that it may be intercepted by unauthorized persons.

6. Advise personnel of the disciplinary action that may result from violation of security regulations.

To achieve maximum benefits from a security education program, the program must cover a variety of security facets as well as being geared to cover a wide cross-section of personnel. It should commence with the indoctrination of new personnel and include periodic refresher training for those personnel who are continually handling classified material. Portions of the program should also cover foreign travel by persons having had access to classified material and the proper debriefing of personnel who are terminating or transferring.

The preceding areas of educational training mentioned actually comprise only the minimum training that should be covered. Additions to these should be made by organizations as their particular missions warrant.

OpNav Instruction 5510.0 (Series) and DOD Regulation 5200.1 (Series) provide guidance for the formation of a sound and complete security education program.

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